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CEMENTATION OF SOME CALIFORNIA TERTIARY RESERVOIR SANDS[1]

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ABSTRACT

Forty core samples of feldspathic sandstones from wells in central and southern California contain the following secondary (authigenic) minerals deposited in the open pores or formed within the detrital clay matrix: quartz, albite, orthoclase, microcline, dolomite, calcite, anatase, kaolinite (?), glauconite, barite, and pyrite. These minerals, together with the primary clay matrix prominent in some samples, are the cementing substances.

Secondary quartz occurs as crystallographically continuous outgrowths on detrital grains. The secondary feldspars are both outgrowths on detrital feldspars and idiomorphic authigenic crystals in cavities. Dolomite takes the form of small rhombs and is most abundant in the clay matrix, whereas calcite is typically an anhedral cement filling open pores. Kaolinite (?) occurs as radial fibrous or granular aggregates in the detrital clay matrix. Dolomite and secondary feldspars have idiomorphic outlines in contact with secondary quartz and calcite, and in many cases calcite has grown on top of secondary idiomorphic quartz.

Intricately penetrating contacts between detrital grains of quartz and feldspar may be the result of grain-on-grain pressure and consequent solution at points of contact between grains. The crystallization of secondary quartz and feldspar may be complementary to such local solution.

INTRODUCTION

This paper reports the first results of an investigation of the characteristics and sequence of deposition of cementing materials and the resulting rock texture in some California Tertiary reservoir sandstones. The results of somewhat similar but more extensive study of Rocky Mountain sandstones made by W. A. Waldschmidt (1941) led the writer to undertake an examination of cements in California sandstones.

This study was limited to forty core samples from four wells of the Standard Oil Company in central and southern California. The samples studied (ten from each of the four wells) represent small stratigraphic intervals in each of the wells, as follows (see also table 2):

Well A.—Pleasant Valley Field, Fresno County, California. The samples from this well are numbered S3-1A to 10A and were taken from depths between 9,069 and 9,142 feet, representing the Gatchell sand of Middle Eocene age.

Well B.—Coalinga Extension Field, Fresno County, California. Samples from this well are S3-1B to 10B, were taken from depths between 8,172 and 8,230 feet, and represent the Gatchell sand of Middle Eocene age.

Well C.—Kettleman Middle Dome, Kings County, California. Samples from this well are numbered S3-1C to 10C,

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[1] Manuscript received July 12, 1948.

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were taken from depths between 12,195 and 12,240 feet, and represent the McAdams sand of Middle Eocene age.

Well F.—Los Angeles Basin north of West Coyote Field, Los Angeles County, California. Samples from this well are numbered S3-1F to 10F, were taken from depths between 10,358 and 10,805 feet, and represent the Puente sandstone of Upper Miocene age.

Thin sections of each sample mounted on a universal stage were investigated by techniques developed during the course of this study but described elsewhere (Gilbert and Turner, 1948). The universal stage proved invaluable and is recommended for sedimentary petrographic examinations, especially for detailed studies of textures and authigenic minerals in thin sections or permanent grain mounts.

The samples were small cores approximately 1 inch in diameter, embedded in wax, on which permeability measurements had been made. The orientation of the cores with respect to the bedding is unknown, but thin sections cut across these cores must be inclined at a considerable angle to the bedding because the thin dimension of the mica flakes was invariably seen in section.

Two of the three formations studied, the Gatchell and McAdams sands, productive zones of Eocene age, are described from the subsurface in the Kettleman Hills–Coalinga area but are not recognized in surface outcrops. They are generally rather well-sorted, permeable, medium-grained sandstones reaching a thickness of several hundred feet in some places but typically lensing so that they form a type of stratigraphic trap. The Puente formation is a thick series of shales and sandstones of Upper Miocene age cropping out in the northern part of the Los Angeles Basin. Although the formation is mostly shale, considerable sandstone occurs in its middle and upper part. In the cores examined, the Puente sandstone, unlike the better-sorted Gatchell and McAdams sands, is characterized by a clay matrix.

MINERALOGY OF THE SANDSTONES

DETRITAL MINERALS

All the sandstones are light colored and feldspathic. Those from wells A, B, and C (Gatchell and McAdams sands) contain on the average approximately 65–70 per cent quartz and 30–35 per cent feldspar most of which is orthoclase or microcline; the Puente sandstone from well F is more feldspathic, containing at least 50 per cent feldspar of which about half is plagioclase. Clay is abundant in some samples but is lacking in others. Detrital heavy minerals make up approximately 1 per cent of the sandstones; they include zircon, tourmaline, garnet, apatite, and sphene with minor amounts of hypersthene, augite, and hornblende. Solution etching appears on grains of garnet, hypersthene, and augite.

SECONDARY MINERALS

Minerals that have crystallized in the sands after their deposition are classed collectively as secondary or authigenic. In the sandstones examined such minerals include quartz, orthoclase, microcline, albite, dolomite, calcite, kaolinite(?), glauconite, anatase, barite, and pyrite. They may have been formed by processes operating at the time of deposition or by chemical activity of connate waters following deep burial. The writer believes the latter origin to be more probable. Possibly some of the secondary minerals crystallized under the influence of regional stress or circulating ground waters of surface origin, but there is no reason to suspect that any of them is the direct product of surface weathering or of igneous hydrothermal emanations.

Classified according to type of cement, the samples may be divided into two groups—one in which the cement is chiefly a clay matrix and another in which chemically precipitated secondary minerals are the chief binding material. Such a subdivision probably could be applied to any extensive suite of sandstones (Waldschmidt, 1941, p. 1857; Krynine, 1948, p. 138). Very fine detrital material, called clay although its specific components are not determinable, fills the spaces between sand grains in many of the samples and by compaction and partial recrystallization to kaolinite has become coherent, forming a cement. Sandstones cemented chiefly by clay include some in which detrital clay and recrystallized kaolinite are the only binding materials and others in which rhombohedrons of dolomite occur throughout the clay and here and there fill the spaces between sand grains. In such rocks calcite and secondary quartz and feldspars are absent or occur in minor amounts—a fact suggesting that the clay has inhibited the normal precipitation of secondary minerals in open pores.

Where detrital clay is lacking or present in only small amounts, chemically precipitated secondary minerals are typically more abundant. In a few samples calcite fills the spaces between sand grains and is the chief cement; in others the cement is secondary quartz, commonly accompanied by less abundant secondary feldspars; and in some quartz, feldspar, and calcite all combine to bind the sand grains together.

Quartz.—Secondary quartz in the form of crystallographically continuous outgrowths on detrital grains is present in every sample from wells A and C and as smaller outgrowths in most of the samples from wells B and F. Where a prominent detrital clay matrix is present, quartz outgrowths are either small or lacking altogether. In samples from wells A and C secondary quartz is the chief cement, but all of it occurs as enlargements of detrital grains, and none in the form of granular aggregates between detrital grains has been found.

Evidence of quartz enlargement is of several kinds. Most convincing are the perfect crystal forms projecting into partly filled cavities. The faces formed cannot always be seen unless the thin section is tilted on a universal stage (Gilbert and Turner, 1948). Even a slight tilt may bring into sharp focus crystal faces which might otherwise be obscure or overlooked entirely. Faces are evident on many of the grains when they are disaggregated and examined with a binocular, but under these circumstances the textural relations are not evident. Some outgrowths can be recognized in thin sections and distinguished from the detrital portion of the quartz by the presence of a dusty line or zone of inclusions representing the original surface of the detrital grain. But in the sandstones studied many, if not most, of the angular detrital grains were originally bounded by clean surfaces so that the contact with the quartz overgrowth may be obscure or invisible. Where the idiomorphic form of the quartz outgrowth, together with a dusty line representing the original surface of the detrital grain is seen, both the existence and the size of the outgrowths are established. The sandstones most tightly cemented by secondary quartz have an interlocking texture in which the added quartz has completely filled much of the original pore space and is molded by the adjacent grains. In such rocks, idiomorphic forms are lacking and the intimately interlocking grain contacts alone are evidence of quartz outgrowth.

Measurements made with a universal stage (Gilbert and Turner, 1948) on many idiomorphic outgrowths show that the faces developed belong to the most common crystal forms of quartz, that is, the unit prism and the positive and negative unit rhombohedrons—$\{10\bar{1}0\}$, $\{10\bar{1}1\}$, $\{01\bar{1}1\}$. All three of these forms are represented on the large outgrowth seen on the central quartz grain of plate 4, *E* and figure 4. Among all the faces determined, a single pyramid, $\{11\bar{2}1\}$, was found on one crystal. This may mean that only the more common forms tend to develop, or it may be that other faces are smaller and cut only rarely by the thin section.

In a number of cases quartz enlargements have been observed on as many as three or four detrital grains surrounding the same void, so that apparently the outgrowths are not added in any one favored direction in the fabric. Where outgrowths from two sides of a cavity have made contact, the bounding surfaces between them commonly appear to be planes, but in no case have they proved to be rational crystal faces of either quartz grain.

Feldspar.—Secondary feldspar includes orthoclase, microcline, and albite and is not nearly so abundant as secondary quartz. However, it has been observed in every sample from well C and in some samples from each of the other wells, and its presence is suspected in yet other samples where positive recognition is not possible. It is distinguished from detrital feldspar primarily by the presence of idiomorphic forms against cavities in the rock together with its water-clear appearance. Both orthoclase and microcline have developed as crystallographically continuous outgrowths on detrital grains of potash feldspar and also in open cavities as idiomorphic crystals not continuous with any detrital grain visible in the thin section. Albite occurs as idiomorphic crystals in voids (pl. 1, *A*), but no albite outgrowths on detrital feldspar have been recognized. All three varieties of secondary feldspar have been observed in the same thin section.

Determinations of the feldspars were made in thin section with the universal stage. The indicatrix and poles of visible crystal faces, cleavage planes, and twinning planes were plotted on equal-area projections and determinations made as described elsewhere (Gilbert and Turner, 1948). The faces most commonly developed are the basal and side pinacoids, $\{001\}$ and $\{010\}$; the unit prisms, $\{110\}$ and $\{1\bar{1}0\}$, the pyramid, $\{\bar{1}11\}$, and dome, $\{\bar{1}01\}$, are common; whereas the prisms, $\{130\}$ and $\{1\bar{3}0\}$, and dome, $\{\bar{2}01\}$, have been observed on only a few crystals. Twinning according to the albite law is visible in some of the larger crystals of secondary albite and microcline but is uncertain or lacking in small crystals. Where detrital microcline is rimmed by secondary microcline, the twinning commonly continues across the boundary into the secondary rim.

There can be no doubt that water-clear idiomorphic feldspar outgrowths attached to cloudy detrital grains and equally clear euhedral crystals in open cavities are both secondary; but there are other water-clear feldspars, apparently uniform throughout and having tight interlocking contacts with surrounding detrital grains of all kinds, which may also be secondary. By careful examination and tilting the stage to appropriate angles, one or two small crystal faces can be observed on a few of these latter crystals where minute voids occur along their contacts; others have only their water-clear appearance to suggest their secondary origin. The writer is not prepared to classify such anhedral feldspars as definitely secondary, but it seems likely that they are. If feldspar has crystallized in pores of the sandstones, as is certain, some such pores may well have been filled completely by water-clear anhedral feldspar molded against the surrounding detrital grains.

Dolomite.—Dolomite is present in most of the samples from well B and in some from well F. Commonly it occurs as scattered rhomb-shaped crystals in the clay matrix or in voids, but here and there granular aggregates of the rhomboidal grains completely fill the pore space between detrital grains. Although its occurrence is not confined to the clay matrix, dolomite rhombs are much more numerous in the clay than in open pores. Apparently the clay matrix afforded a more favorable environment for the crystallization of dolomite than existed elsewhere in the rock; likewise dolomite tends to form in altered biotite more commonly than in open pores (pls. 3, *E*, and 4, *C* and *D*). However, there is no constant association of these minerals, and many samples containing both detrital clay and abundant biotite do not contain dolomite.

The writer would hesitate to say that dolomite never occurs in these sandstones as anhedral grains; but wherever positive identification has been made, this mineral occurs in idiomorphic or partly idiomorphic crystals bounded by the unit rhombohedron $\{10\bar{1}1\}$. Identification of the mineral as distinct from calcite was made in grain separates on the basis of specific gravity and refractive index and in thin section by measurement with the universal stage of the angle between the optic axis and the direction of a ray (e') having the index of refraction of adjacent quartz or balsam (Gilbert and Turner, 1948). After experience had shown that the idiomorphic carbonate was invariably dolomite, the idiomorphic habit was used as a basis for identification.

Calcite.—With the exception of samples from well A, calcite occurs in most of the sandstones studied, but its amount varies greatly. In some samples it is the only cement and completely fills the pore spaces between the detrital grains (pl. 1, *C*), and in such cases it is either medium granular or composed of very large crystals each uniformly oriented over an area sufficient to enclose many detrital grains. Elsewhere, only a few calcite grains, invariably anhedral and usually filling two or three adjacent pores, are present in a slide (pl. 2, *A*, and fig. 2). There is no constant association of calcite with dolomite; the calcite occurs both in those sandstones devoid of clay and in those containing abundant clay and, unlike dolomite, has not formed in the clay matrix but is restricted to what were apparently open pores in the original rock. No idiomorphic forms have been observed on calcite, but lamellae parallel to the flat rhombohedron $\{01\bar{1}2\}$ were commonly seen.

Some detrital grains of quartz and feldspar have been penetrated by calcite—a relation that can usually be correlated with crystallization of the carbonate along fractures or cleavage in the detrital grains. In some such cases the calcite has wedged apart the sand grains (pl. 1, *B*). Replacement of quartz or feldspar by carbonate, either dolomite or calcite, was not observed and is certainly not typical in the sandstones studied.

Kaolinite.—Partial recrystallization of the detrital clay is common, and the product is either a granular aggregate or several large crystals, commonly showing a radial structure where the cleavage is inclined at a high angle to the plane of the thin section. Tentatively, this recrystallized clay has been identified as kaolinite. It is biaxial and negative with a medium optic axial angle, and the acute bisectrix (X) is essentially normal to the single cleavage. The birefringence is low, and the refractive index of the fast ray (α) is a little greater than the high refractive index of quartz. Its constant association with the detrital clay matrix suggests that this is a reasonable determination.

That the kaolinite is indeed a secondary mineral seems certain. The largest grains of it, particularly those showing radial structure, are so oriented and molded around detrital grains of quartz and feldspar as to preclude any suggestion that they, too, are detrital. Nor do they show any of the parallel orientation or distortion so typical of the detrital mica in these rocks.

Anatase.—Although it is not abundant enough to constitute an important cement, anatase is a secondary mineral characteristic of this suite of sandstones. In some samples it forms as much as 1 per cent by weight of the rock. It is readily recognized as idiomorphic crystals typically tabular parallel to the basal plane {001} but occasionally pyramidal and having a yellowish color. Single euhedral crystals line some of the voids in the rocks (pl. 1, *D* and *E;* fig. 1), and cloudy aggregates between detrital grains constitute a local cement. The observations made have not certainly proved any particular mode of origin for the anatase; but because a little sphene is present in heavy residues from some of the samples, it may be assumed that anatase has formed from the recrystallization of the titania present in sphene. Also a few crystals present in altered biotite suggest that some of the anatase may have formed from the recrystalliza-

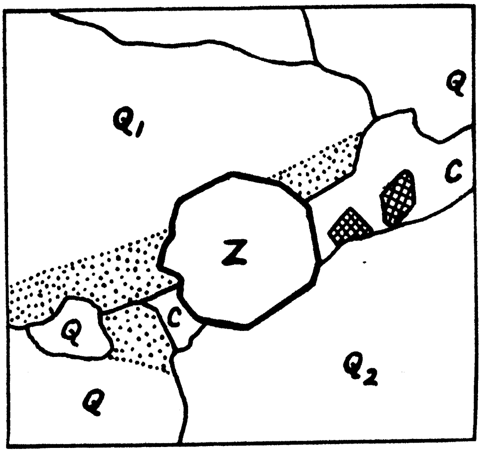


FIG. 1.—Tracing of subidiomorphic zircon (Z) partly embedded in detrital quartz (Q_1 and Q_2) and partly surrounded by secondary quartz outgrowths (*stippled*). Cross-hatched grains are anatase crystals in cavity (C).

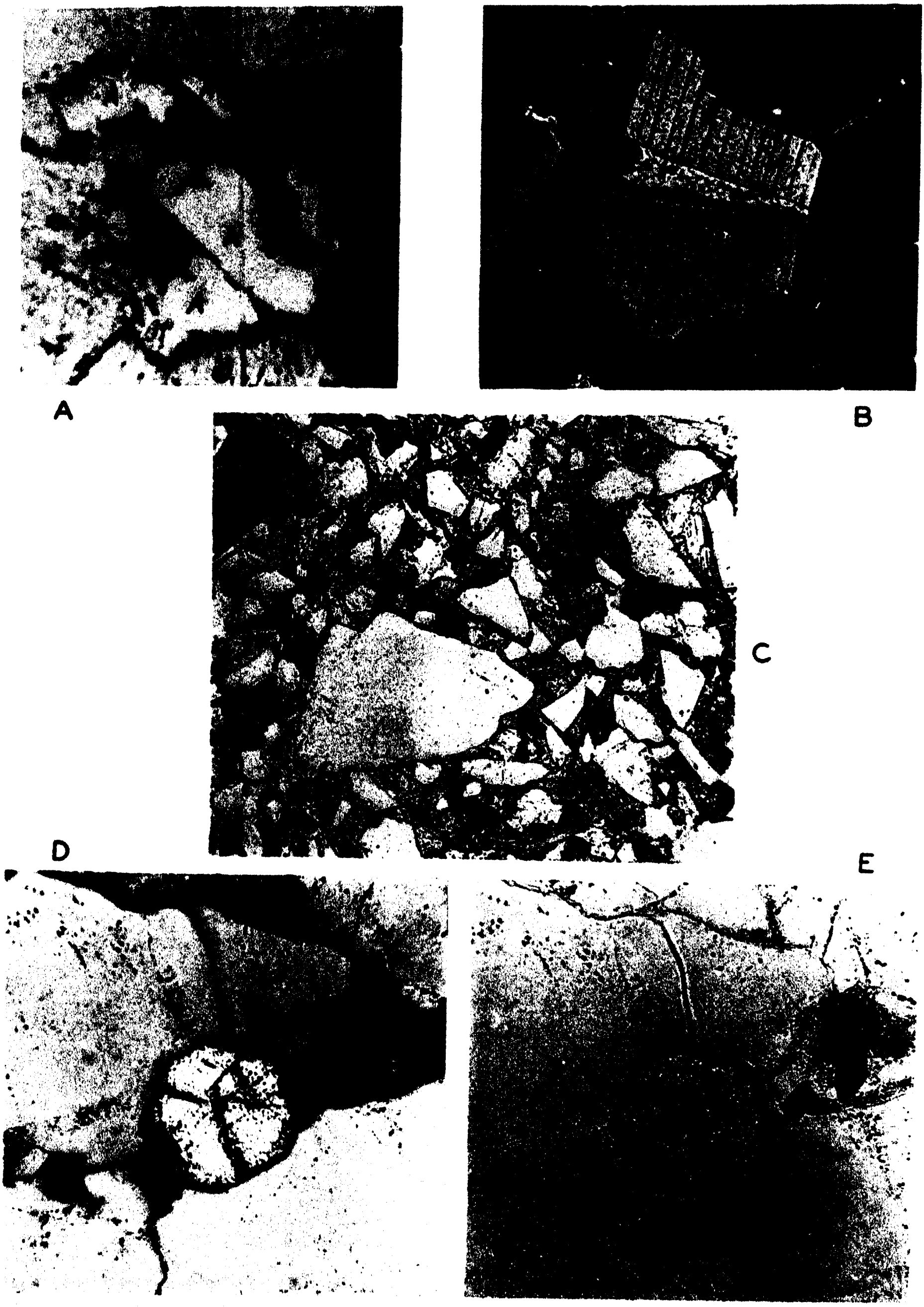
PLATE 1

A, Photomicrograph of idiomorphic albite (*A*) developed in pore space between detrital grains of quartz (*Q*), orthoclase (*O*), and plagioclase (*P*). Photograph taken under crossed nicols with the universal stage tilted about 25°, with the result that only the central albite crystal is in focus and shows its habit. McAdams sand (S3-10C); magnification 275×.

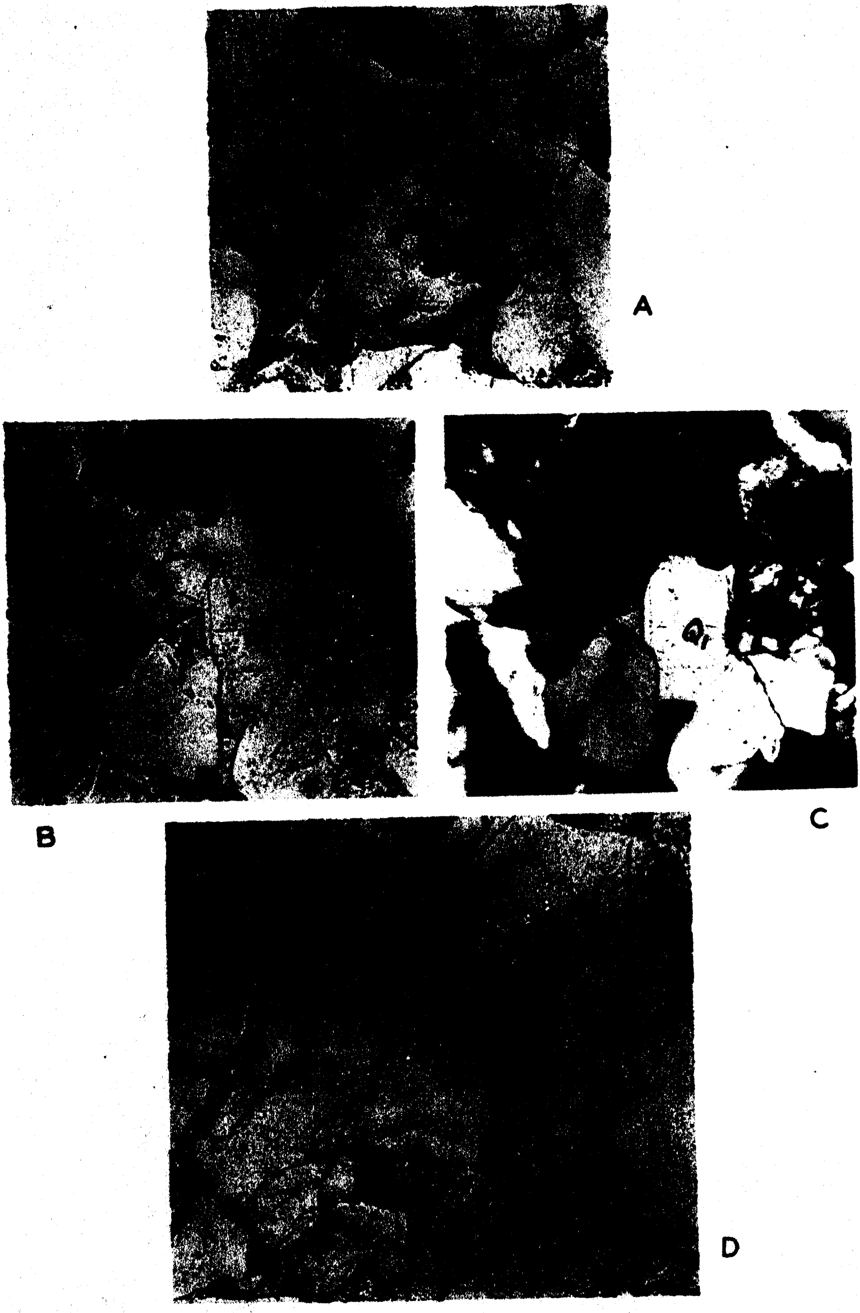
B, Photomicrograph of detrital plagioclase (Ab_{90}) surrounded by detrital clay matrix and wedged apart by secondary carbonate crystallized along the basal cleavage. Puente sandstone (S3-3F); magnification 250×.

C, Photomicrograph of loosely packed sand grains (mostly quartz and feldspar) cemented by a continuous calcite matrix. Gatchell sand (S3-10B); magnification 50×.

D and *E*, Photomicrographs (*D*, crossed nicols; *E*, plane polarized light) of subidiomorphic zircon partly embedded in detrital quartz and partly surrounded by secondary quartz outgrowths. McAdams sand (S3-5C); magnification 175×. See also fig. 1.



Cementation of reservoir sands



Cementation of reservoir sands

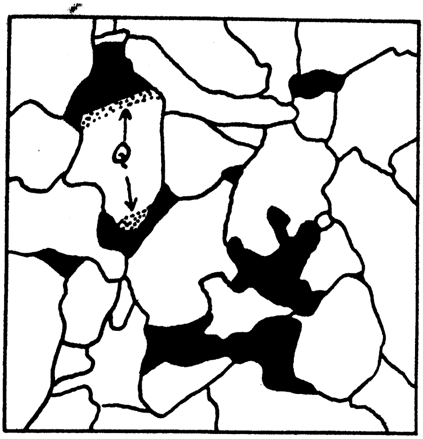


FIG. 2.—Tracing illustrating local calcite and also the interlocking relations of detrital quartz and feldspar in the McAdams sand (S3 = 10C). Calcite (*black*) is uniformly oriented throughout and is molded over small quartz outgrowths (*Q*, *stippled*).

tion of titanium oxide in biotite. No other titanium-bearing minerals have been noted.

Barite.—Barite was found only in the shallowest sample from well A where it is a local cement, occurring as a single optically continuous grain that fills the original pores throughout a considerable portion of a thin section (pl. 2, *D*).

Pyrite.—The mineral identified as pyrite is opaque, fine grained, and has the usual color of pyrite, but it could conceivably be marcasite. It is not abundant in any sample but occurs in very small amounts in all of them. Typically it forms tiny microcrystalline aggregates (microconcretions) less than 100 microns in maximum diameter, but its occurrence as a local cement in one thin section was noted.

Glauconite.—A few grains of this mineral occur in the clay matrix in samples S3-5A, 6A, 2B, and 3B. It has the usual cloudy green appearance typical of the mineral in thin section.

MUTUAL RELATIONS BETWEEN SECONDARY MINERALS

General statement.—Where, as is commonly the case, several secondary minerals can be observed in a single thin section, their mutual relations, except in the cases of barite and pyrite, have been carefully examined. In many sections one mineral is constantly idiomorphic with respect to another—a relation capable of at least three alternative interpretations. It could conceivably arise by simultaneous crystallization (as in metamorphism) under conditions such that one mineral has a greater tendency to assume its own crystal outline than the other; or it could reflect a time sequence of crystallization, in which the idiomorphic mineral crystallized first; or possibly the idiomorphic mineral is the later and has replaced other minerals (Krynine, 1948, fig. 12). Available evidence is still inconclusive except for cal-

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PLATE 2

A, Photomicrograph illustrating local calcite cement and also the interlocking relations of detrital quartz and feldspar in the McAdams sand (S3-10C). Calcite is uniformly oriented throughout and is molded over small quartz outgrowths (*Q*). Magnification 80×. See also fig. 2.

B and *C*, Photomicrographs (taken with plane polarized light and crossed nicols) illustrating (1) secondary quartz outgrowth molded around idiomorphic orthoclase (*F*) and (2) calcite (*arrow*) crystallized on top of idiomorphic quartz outgrowths. The clear orthoclase (*F*) is bounded by the crystal forms {110}, {001}, and {101} and is surrounded by an outgrowth of detrital quartz (Q_1). The crystal faces bounding the triangle of calcite in the center of the photograph are rhombohedron faces on quartz outgrowths from detrital grains Q_1 and Q_2 and are not rational faces of the calcite. McAdams sand (S3-7C); magnification 90×.

D, Photomicrograph of barite cementing detrital quartz grains in the Gatchell sand (S3-1A). The mineral having high relief is barite and is in optical continuity throughout the area of the photograph. Magnification 40×.

cite, some of which is undoubtedly later than the other secondary minerals. Calcite not only is consistently anhedral but occupies the central portions of cavities whose margins are lined by other secondary minerals.

The observed relations suggest the following general conclusions for the sandstones studied:

1. Calcite is definitely later than secondary quartz and feldspars; but in a few samples (pl. 1, *C*, for example), where calcite is the only significant cement, it may have crystallized early and inhibited the development of any other secondary minerals.

2. Some secondary feldspar crystallized simultaneously with secondary quartz although some probably crystallized before the quartz.

3. Dolomite crystallized before secondary quartz and calcite.

4. Anatase crystallized before most of the calcite and probably before the quartz outgrowths.

5. Kaolinite, formed by recrystallization of detrital clay, probably formed after deep burial during the same general period when many of the other secondary minerals crystallized, but its relations to them are obscure.

6. Pyrite and barite are thought to have crystallized early among the secondary minerals.

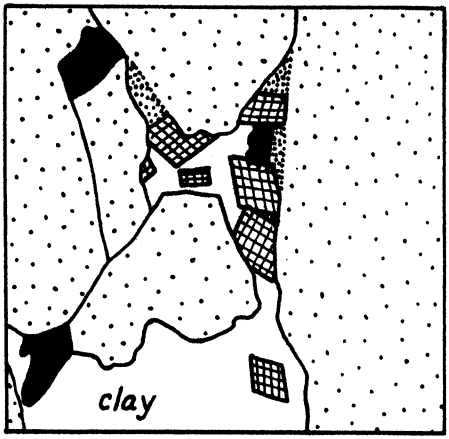

FIG. 3.—Tracing slightly idealized from plate 3, *A*, to indicate the relations determined by tilting to various angles on a universal stage but not clearly evident at any one angle of tilt as in the photographs. Dolomite rhombs (*cross-hatched*) are molded against detrital quartz (*wide stippling*); quartz outgrowths (*close stippling*) are molded against rhombohedron faces of several dolomite crystals. Blank areas in the sketch are detrital clay; black areas are cavities.

The secondary minerals listed above include some not observed by Waldschmidt (1941, p. 1858) in sandstones of the Rocky Mountain region; but quartz, dolomite, and calcite were listed by him as having formed in the order given. The author's results, therefore, differ from Waldschmidt's in that the relative order of crystallization of quartz and dolomite is reversed and also in that some of the calcite in the California sandstones may

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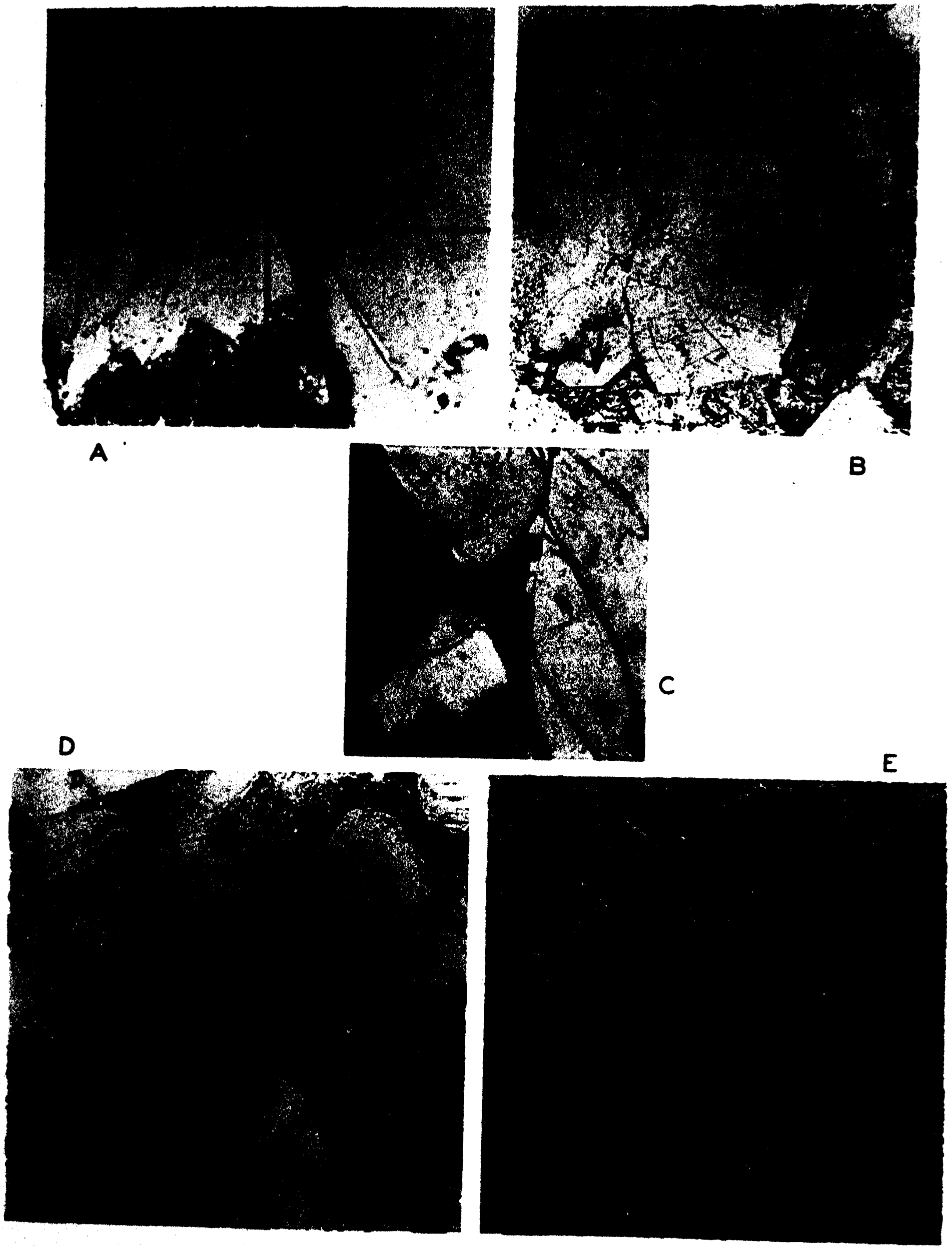
PLATE 3

A and *C*, Photomicrographs illustrating the relationship between dolomite and secondary quartz outgrowths in Puente sandstone (S3-1F). Secondary quartz outgrowths are clearly evident in photograph *C* taken under crossed nicols. Magnification of *A*, 110×. See also fig. 3.

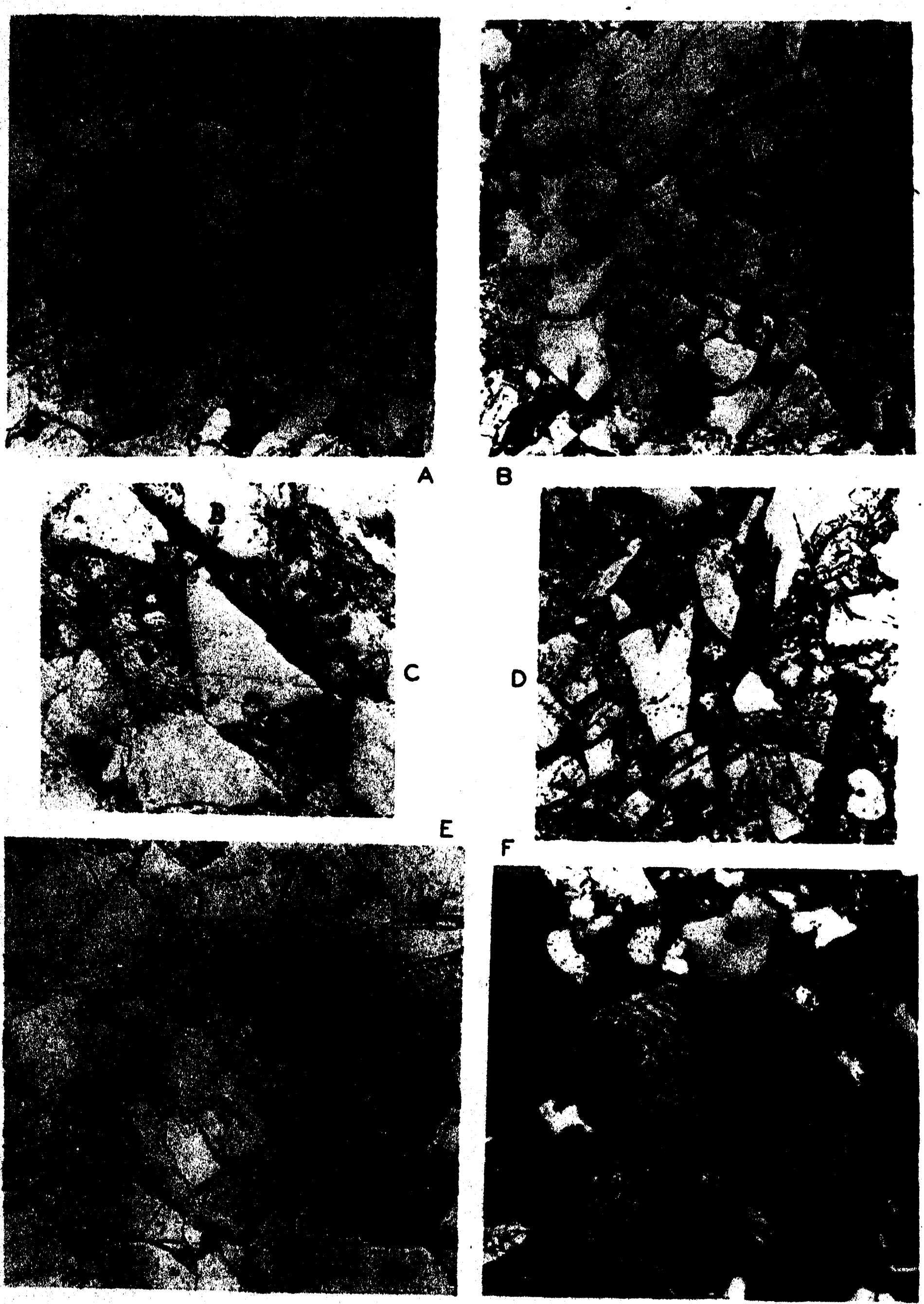
B, Photomicrograph of calcite (*high relief*) surrounding detrital quartz grains with secondary outgrowths (*arrows*) in McAdams sand (S3-1C). The outgrowth in the upper left actually has three crystal faces visible, but only one is apparent at the angle of tilt used for the photograph. Magnification 52×.

D, Photomicrograph of well-sorted Gatchell sand (S3-9B) with open pores shown in black. Detrital grains of quartz and feldspar tightly packed but not typically interlocking (compare pl. 4, *A* and *E*) and very slightly cemented. Incipient outgrowths of quartz (*Q*) and orthoclase (*F*) are shown by arrows; two small dolomite crystals (*D*) in open pores are also indicated. This is the most permeable sample studied and is as porous as any. Magnification 40×.

E, Scattered rhombs and granular aggregates of dolomite in detrital clay matrix of Gatchell sand (S3-2B). Magnification 40×.



Cementation of reservoir sands



Cementation of reservoir sands

be somewhat earlier than secondary quartz.

Quartz and feldspars.—Secondary quartz and feldspars occur together in all samples from well C (McAdams sand) and in some samples from other wells. In every case the relation between them is the same, the feldspar being idiomorphic in contact with the quartz. This relation applies alike to orthoclase, microcline, and albite against quartz. As previously indicated, however, the writer considers that the periods of crystallization of these two minerals—quartz and feldspar—have overlapped somewhat although the feldspar perhaps tends to be the earlier.

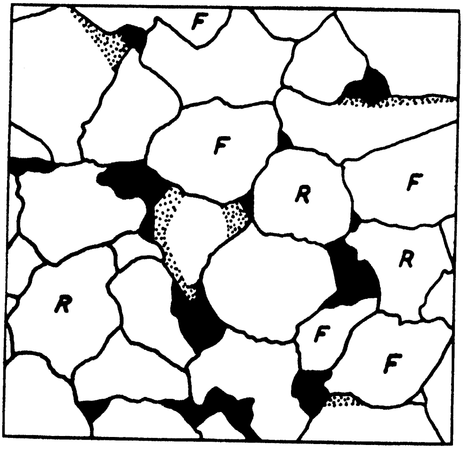


FIG. 4.—Tracing showing idiomorphic quartz outgrowths and tightly interlocking texture of detrital grains in McAdams sand (S3-3C). Unlabeled grains are detrital quartz, stippled areas are quartz outgrowths, black areas are cavities; *F* represents detrital feldspar, and *R* indicates fine-grained rock fragments.

The relation between secondary quartz and albite crystallized in the same cavity is illustrated by a thin section of the deepest sample from well C (S3-10C). Plate 1, *A*, is a photograph of the albite in this cavity taken on a universal stage with a tilt so high that only the albite is in focus. Several pure albite crystals having Baveno habit project into the cavity, and none is in crystallographic continuity with any detrital grain visible in the thin section. The crystals appear to have crystallized on top of thin outgrowths of secondary quartz, as if a little quartz had crystallized ahead of the albite, but the quartz outgrowths have also partly surrounded the albite crystals and are molded against well-formed faces of the feldspar (Gilbert and Turner, example 4). This relation is interpreted as the result of essentially simultaneous crystallization of quartz and albite.

A relationship between secondary quartz and orthoclase is illustrated in plate 2, *B* and *C*, where a distinct out-

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PLATE 4

A, Photomicrograph illustrating the interlocking texture of detrital quartz and feldspar grains and showing one greatly distorted crystal of detrital biotite (*arrows*) in McAdams sand (S3-8C). Magnification 42×.

B, Photomicrograph of detrital biotite (*arrows*) distorted by the tight packing of Puente sandstone (S3-9F). The texture is not typically interlocking, and considerable detrital clay is present. Magnification 42×.

C, Dolomite rhombs replacing biotite (*B*) and also embedded in calcite cement (*C*). Gatchell sand (S3-4B); magnification 58×.

D, Dolomite rhombs embedded in continuous calcite cement. Gatchell sand (S3-10B); magnification 42×.

E, Photomicrograph showing idiomorphic quartz outgrowths and tightly interlocking texture of detrital grains in McAdams sand (S3-3C). Crystal faces on the outgrowths from the central quartz grain include the prism and both rhombohedron forms. Magnification 54×. See also fig. 4.

F, Photomicrograph (taken under crossed nicols) of distorted plagioclase in Puente sandstone (S3-5F) having a detrital clay matrix. Strain and radial and concentric fractures in plagioclase (P_1) appear to center around penetrating point of quartz grain (*Q*); twin lamellae bent and radial cracks developed in plagioclase (P_2). Magnification 75×.

growth of quartz is molded against three well-developed faces—the forms $\{001\}$, $\{1\overset{\pm}{1}0\}$, and $\{\bar{1}01\}$—on a crystal of orthoclase projecting into a void. Where the quartz outgrowth is in contact with calcite and projects into the unfilled portion of the cavity, typical prism and rhombohedron faces of quartz are clearly seen.

Quartz and calcite.—These two minerals, both secondary, are commonly in contact. Typically the calcite is anhedral and molded against the idiomorphic forms of quartz outgrowths (pls. 2*A*, *B*, *C;* 3, *B;* and fig. 2), and the only reasonable interpretation of this relation is that the quartz crystallized first. The detrital grains were enlarged and the open pores of the sandstone thus partly filled by secondary quartz; later calcite crystallized in some of the remaining voids and filled them completely. Such a relationship has been observed many times in these sandstones. However, it is not always demonstrable by means of ordinary microscopic technique, for proof requires that the crystal faces forming the boundaries between quartz and calcite be tilted into sharp focus and identified as belonging certainly to one or the other of the two minerals. This can readily be done with a universal stage. A pertinent example is illustrated in plate 2, *B* and *C*, where what appears to be a rhombohedron of calcite is surrounded by quartz outgrowths from two detrital quartz grains. This "rhombohedron" proves, however, to be bounded by rational faces (the form $\{10\bar{1}1\}$) of the quartz outgrowths having no relation whatsoever to normal faces of calcite (Gilbert and Turner, 1948, example 2).

In certain other rocks in which calcite is very abundant and fills the space between detrital grains, the calcite may have crystallized earlier. In such rocks secondary quartz outgrowths have not been found although they are present in closely adjacent samples of the same formation where calcite is less abundant. This suggests that locally where calcite is very abundant it crystallized early, completely cemented the rock, and inhibited the growth of other secondary minerals. Further evidence is found in samples like that illustrated in plate 1, *C*, where detrital grains in a continuous calcite cement do not touch one another, in contrast to the close packing of detrital grains in similar samples not so completely cemented (pl. 3, *D*). This may indicate that the abundant calcite cement crystallized early before tight packing of the detrital grains occurred.

A single case (sample S3-4B) where secondary quartz may have replaced calcite has been observed. The calcite surrounds an idiomorphic quartz crystal greatly elongated parallel to one of the rhombohedrons, instead of the usual prism, and is strictly parallel to one of the cleavages of calcite. Admittedly, as an alternative to the suggested replacement of calcite by quartz, this relationship may be interpreted as calcite molded against a quartz crystal of abnormal habit with the rhombohedrons of the two minerals in parallel orientation.

Quartz and dolomite.—Although dolomite and quartz outgrowths commonly occur in the same sample, they have rarely been observed in contact. Plate 3, *A* and *C*, and figure 3 illustrate the mutual relationships of dolomite and secondary quartz. Dolomite rhombohedrons in a void containing some clay are molded against detrital quartz grains around the cavity, whereas a secondary outgrowth of quartz on one side of the cavity is molded against the rhombohedron faces of dolomite as though it had crystallized after the carbonate. This relationship, like that between quartz and feldspar, is subject to alternative interpretations; either the dolomite crystallized ahead of the secondary quartz or the two minerals crystallized simultaneously. The writer favors the first of these interpretations, especially in view of the observed relationships between dolomite and calcite which suggest that dolomite may have formed even before the early calcite. A third alternative, namely, that the dolomite replaced the secondary quartz, is untenable; if this had happened, the dolomite should show euhedral boundaries against the detrital quartz as well as against the quartz outgrowths. Indeed, dolomite in contact with detrital quartz and feldspar is invariably molded against the surfaces of those grains; although what appear at first to be penetrations of detrital quartz by idiomorphic dolomite have been observed in many samples, tilting of the stage has shown in every case that the actual relationship is one in which the dolomite is simply attached to the surfaces of the detrital grains.

Dolomite and calcite.—Wherever these two minerals are in contact, idiomorphic dolomite rhombohedrons are embedded in anhedral calcite cement. This relationship is illustrated in plate 4, *C* and *D*. One of these examples (pl. 4, *D*) is from a sample (S3-10B) in which calcite is very abundant and probably had crystallized early. Clearly, the dolomite rhombohedrons in calcite may be explained either as a replacement of calcite by dolomite or, alternatively, as dolomite crystals formed early and later surrounded by calcite cement. The evidence is inconclusive. The dolomite rhombs have no observed consistent orientation with respect to the enclosing calcite, as might be anticipated in the event of the suggested replacement. Furthermore, whether associated with calcite or occurring in clay-filled cavities or in association with biotite in rocks free of calcite, dolomite occurs in part as idiomorphic rhombs attached to detrital quartz and feldspar and also as apparently perfect isolated crystals. The mutual relationships of dolomite and calcite with respect to secondary quartz indicate that dolomite formed ahead of the calcite in most samples.

Dolomite and kaolinite.—The secondary clay mineral, thought to be kaolinite, here and there contains rhombs of dolomite. In one case the cleavage of the kaolinite appeared bent around the dolomite crystals as though the carbonate had crystallized later and made space for itself by physical distortion of the kaolinite; in other cases no such relationship was observed, and there is no indication as to which mineral is the earlier. Dolomite, which tends to occur most commonly in the clay matrix, may have replaced the kaolinite itself or it may have replaced the detrital clay matrix before recrystallization produced the kaolinite.

Anatase.—Most of the anatase seen in thin sections is in aggregates of minute crystals clustered together in cavities; less commonly it occurs, as in plate 1, *D* and *E*, figure 1, in larger individual crystals along the margins of the cavities. Generally its relationship to other secondary minerals is not evident; but in several cases anatase crystals attached to detrital quartz grains are surrounded by and probably are earlier than calcite. Because a few minute crystals have been surrounded by quartz outgrowths, the anatase was probably formed early among the secondary minerals.

PREFERRED ORIENTATION

In order to determine whether there is any preferred orientation of quartz in the most tightly packed and cemented sandstones, the positions of the optic axes in three hundred quartz grains in each of several thin sections were determined with a universal stage. Plotted on an equal-area projection net, a random orientation is indicated by all these measurements. Figure 5 is a typical result.

Measurements of the orientation of optic axes in dolomite rhombohedra likewise indicate a random orientation in those few slides that contain abundant dolomite. Detrital mica, however, tends to be oriented parallel to the bedding, as would be expected.

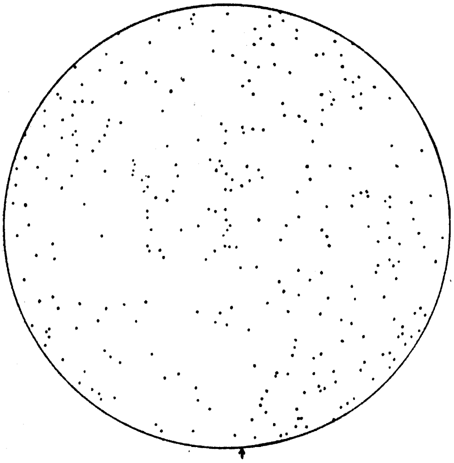

FIG. 5.—Equal-area projection of optic axes in 300 quartz grains (sample S3-3C).

TEXTURE OF THE SANDSTONES

CLASSIFICATION

Typically the sandstones examined are tightly packed; that is, the sand grains are arranged with respect to one another so that minimum bulk volume and porosity is approached. That this tight packing is the result of grain adjustments made under the pressure of deep burial is suggested by the distorted form of detrital biotite grains bent around and squeezed between the more rigid sand grains of quartz and feldspar (pl. 4, *A* and *B*). In some of the sandstones the detrital grains have intimately interlocking and often minutely serrated contacts, which cannot be referred to simple mechanical packing under pressure, but which strongly suggest some partial solution along grain contacts (pl. 4, *A* and *E;* fig. 4). Distortion by fracturing of such interlocked grains, as is illustrated in plate 4, *F*, is not common or typical, but undulatory extinction is observed in many grains. Waldschmidt (1941, pp. 1859–1863) has described and illustrated similarly interlocking quartz grains in the sandstones of the Rocky Mountain region; but in the California sandstones, feldspars and rare minerals like zircon (pl. 1, *D* and *E;* fig. 1), as well as quartz, are involved in the interlocked fabric.

Classified according to texture, the sandstones fall into two broad categories, namely, those in which the sand grains are more or less tightly packed but not typically interlocked and those in which interlocked contacts between detrital grains dominate the fabric. When each of these types is further subdivided according to the character of cementing material, a relation between textural types and character of cement is apparent. Interlocked textures are found in rocks cemented largely by secondary quartz, together with some feldspar, whereas those sandstones without typically interlocked contacts between detrital constituents are uncemented or are cemented by clay either alone or with carbonates. Waldschmidt also noted that interlocked texture was typically associated with relatively large amounts of secondary quartz.

The textural classification given in table 1, based on degree of packing and interlocking of detrital grains and on the nature of cementing material, is satisfactory for the California sandstones and is intended to apply only to those rocks. But it can probably be extended and modified to give a more general classification as the need arises. There are, as might be expected, gradations between the various textural types indicated. Sandstones of class II (not typically interlocked) commonly contain some grains that are mutually interlocking and may also contain small amounts of secondary quartz and feldspar; and those of class III (typically interlocked) may contain grains that are not interlocking. Nevertheless, a division into these classes can be made by brief inspection of thin sections if only the dominant textural mode is sought.

In sandstones, classified as loosely packed, where the detrital grains appear to be separated by a continuous calcite cement (pl. 1, *C*), the apparent separation between sand grains is not large; indeed, when viewed in thin sections cut at random, an apparent separation between many grains is to be expected even though they are in actual contact at some points. The designation "loosely packed" seems appropriate simply by comparison with the more closely packed sand grains in other rocks belonging to textural class II. Very widely separated sand grains more or less evenly distributed through a calcite matrix (sandy limestone), a textural type described and illustrated by Waldschmidt (1941, pl. X and pp. 1863–1866) from the Jurassic and Cretaceous sandstones of the Rocky Mountains, have not been observed in the California sandstones; such rocks probably would not fit into the textural classification given in this paper.

Three well-sorted sandstones (S3-7B, 8B, and 9B), representing several feet of well B, consist of very tightly packed sand grains with very little cement of any kind (pl. 3, *D*). These samples are the basis for the textural class II, A. They are the most porous and permeable samples studied, but the fact that the highest porosity in any of them is but 27 per cent indicates that they are indeed tightly packed. The contrast in texture between these three samples and the loosely packed, calcite cemented sandstone (S3-10B, pl. 1, *C*) 8 feet deeper in the same well is noteworthy because the samples all lie so close together in the same formation and indicates that marked textural differences may be found in a single formation.

TABLE 1

TEXTURAL CLASSIFICATION OF SANDSTONES STUDIED

I. Sand grains rather loosely packed
    A. Calcite cement (pl. 1, *C*)
II. Sand grains tightly packed but not typically interlocked
    A. Poorly cemented (pl. 3, *D*)
    B. Cemented by
        1. Clay (pl. 4, *B*)
        2. Clay and subordinate carbonates
        3. Calcite and subordinate clay
III. Sand grains typically interlocked (pl. 4, *A* and *E;* fig. 4)
    A. Cemented by*
        1. Quartz
        2. Quartz and some calcite
        3. Quartz with a little clay

* Secondary feldspar has been observed in small amounts in sandstones of this textural type and is typically present in textures III, A, 1 and 2.

Other tightly packed sandstones included in textural class II are better cemented, the pore spaces between sand grains being largely filled with clay or with clay and dolomite; a few contain calcite cement and most of them include a small amount of secondary quartz. In contrast, not only are samples from well C and some from well A tightly packed but the sand grains are interlocked. All these samples, classified in textural group III, contain relatively large amounts of secondary quartz and some feldspars, and many of them contain some calcite or clay as well. But the consolidation of these interlocked sandstones has been accomplished not alone by the cementing substances but also by the mutually molded contacts of the *detrital* grains.

ORIGIN OF THE INTERLOCKED FABRIC

The interlocked fabric as a whole is thought to result from differential solution between detrital grains, combined with the complementary outgrowth of secondary minerals in the voids. This is the origin suggested by Waldschmidt (1941, pp. 1859–1863) for a similar texture observed in the quartz sandstones studied by him, and it seems just as applicable in the present case. Differential solution at points of contact between quartz grains in the St. Peter sandstone has also been suggested by Lamar (1928, pp. 47–48) to account for certain pits observed on the surfaces of detrital quartz grains. The typical association of secondary quartz and feldspars with the interlocked fabric strongly supports the idea that these minerals are indigenous to the rock, being formed by partial solution of detrital grains under pressure at points of contact and precipitation of the dissolved substances at places of lower pressure in the voids. Tight, minutely serrated contacts between detrital grains and apparent penetration of one by another without physical distortion are features that can be explained only by postulating some sort of solution along grain contacts, as described recently by Sloss and Feray (1948, pp. 3–14). Certainly, an indigenous source for secondary quartz and feldspars appears more likely than the alternative that quartz and feldspars were precipitated by ground waters which brought the necessary substances in solution from some outside source, a source not connected with igneous activity. The ground waters required would have to be complex solutions of very slightly soluble substances. Their existence, therefore, is very unlikely and, moreover, would not explain why the deposition of the secondary minerals is so closely correlated with solution along the boundaries of the detrital minerals of similar composition.

The existence of clay between the sand grains would tend to prevent or make less effective the mechanism of differential solution and complementary deposition just outlined. Compact clay tends to transmit more uniformly the stress due to overburden and thus to reduce the heterogeneity of the stress and its concentration at points of solid contact between sand grains. Indeed, if sufficient clay is present, the sand grains may not touch each other at all. Thus it is to be expected that interlocked detrital quartz and feldspar together with secondary outgrowths will, in general, be found only here and there in clay-cemented sandstones.

The foregoing outline is only the most general expression of what must be a most complex process involving a great variety of factors about which little is now known. The writer is not prepared to make a detailed analysis of what the process may involve, but enumeration of some of the factors which undoubtedly are involved will serve to emphasize the complexity of the problem. Such are: the general stress condition in the grains as conditioned by the degree of confinement of the pore fluid, the depth of burial, the shapes of individual grains, and the manner in which the grains are packed and touch each other; anisotropy of the minerals composing the individual grains together with the orientation of individual grains with relation to one another and to the directions of applied stress; partial molal volumes of the minerals involved; surface tension effects; relative compressibility of liquid and solid phases; and composition of the pore fluid. Complex though the process may be, the evidence visible in the sandstones examined amply justifies consideration of such a mechanism to account for cementation in many sandstones. It may well be questioned, however, whether the suggested process is effective under a wide variety of conditions and in a wide variety of fragmental rocks deeply buried or whether it is an effective process only under particular conditions and in certain types of sandstone. Future investigation will have to answer that question. For the process to be significantly effective, however, there must be an unequal distribution of stress over the grain surfaces so that solution will occur at points of greatest stress and deposition at places of lesser stress. This condition can probably occur whether or not the pore solution is confined or free to migrate. In any case, the writer envisages the process as being slow and probably as proceding with diminishing effectiveness as the surfaces of contact between grains become larger and larger. It may be that cementation by this method alone is rarely complete. Probably, too, the process would vary in effectiveness with depth of burial, for it is conceived to operate under the pressure induced by loading and under the temperatures to be expected as a result of the normal thermal gradient.

POROSITY AND PERMEABILITY

A relationship between porosity, permeability, and textural and mineralogic type of sandstone is to be expected. Fine grain size and poor sorting (argillaceous sandstone, for example) are well-known syngenetic causes for low permeability; likewise, the postdepositional processes of compaction, cementation, and solution all produce changes in the original porosity and permeability of the deposit. Among the samples examined during the course of this study (table 2) those which have an interlocking texture and are cemented by secondary quartz, feldspar, and some calcite but without any clay (well C) tend to have lower average porosity and permeability than those samples containing compacted clay matrix with or without other cementing materials. The lowest porosity and permeability is shown by a sample loosely packed but solidly cemented with calcite (S3-10B). As might be anticipated, the highest porosity and permeability occurs in sandstones containing little cement and represented by three samples (S3-7B, 8B, 9B) from well B. These results apply only to the sandstones studied. Other sandstones may show greater variations in compaction and cementation.

A general measure of the decrease in porosity and permeability caused by differential solution and complementary deposition, resulting in the interlocked texture, is suggested by this study. The three well-sorted, tightly packed sandstones containing little cement of any kind (S3-7B, 8B, 9B) are comparable in grain size, sorting, and composition with tightly interlocking samples from well C, and they are thought to represent approximately the condition sandstones from well C attained after tight physical packing but before cementation. Thus the difference in porosity and permeability between these two groups of samples (textural classes II, A, and III, A) gives the general order of magnitude of reduction in porosity and permeability in sands such as these, produced by partial solution along grain contacts under pressure and complementary infilling of the

TABLE 2

DEPTH, PERMEABILITY, POROSITY, AND TEXTURE OF THE CORE SAMPLES STUDIED

	Sample No.	Well Depth (Feet)	Permeability (Millidarcys)	Porosity (Per Cent)	Texture* (Table 1)	Average Grain Size†
Well A: Gatchell sand, Middle Eocene	S3-1A	9,069	229	15	II, B, 1	MC
	-2A	9,079	250	14.9	II, B, 1	M
	-3A	9,101	89	13.4	III, A, 3	M
	-4A	9,104	45	12.8	III, A, 3	M
	-5A	9,106	6	12.0	II, B, 1	MF
	-6A	9,112	0.44	9.7	II, B, 1 ?	F
	-7A	9,123	57	12.8	II, B, 1	MC
	-8A	9,133	212	15.4	II, B, 1 ?	MC
	-9A	9,139	29	12.9	III, A, 3 ?	M
	-10A	9,142	88	14.3	III, A, 3	M
Well B: Gatchell sand, Middle Eocene	S3-1B	8,172	99	16.9	II, B, 2	F
	-2B	8,176	84	17.6	II, B, 2	M
	-3B	8,182	240	24.5	II, B, 2	F
	-4B	8,186	20	19.3	II, B, 2	F
	-5B	8,201	254	18.4	II, B, 2	MF
	-6B	8,213	358	19.9	II, A, & II, B, 2	MF
	-7B	8,220	2,020	23.6	II, A	MF
	-8B	8,221	2,180	27.2	II, A	M
	-9B	8,222	2,840	26.9	II, A	MF
	-10B	8,230	0.24	6.7	I, A	F
Well C: McAdams sand, Middle Eocene	S3-1C	12,195	17.8	14.3	III, A, 2	M
	-2C	12,200	3.7	12.2	III, A, 2	MF
	-3C	12,202	9.5	13.1	III, A, 2	M
	-4C	12,207	7.4	11.8	III, A, 2	MF
	-5C	12,221	8.3	12.9	III, A, 2	MF
	-6C	12,224	3.0	12.2	III, A, 2	F
	-7C	12,233	1.7	10.2	III, A, 2	F
	-8C	12,235	2.2	11.3	III, A, 2	F
	-9C	12,237	1.3	10.9	III, A, 1	F
	-10C	12,240	1.7	10.9	III, A, 2	F
Well F: Puente formation, Upper Miocene	S3-1F	10,358	36	16	II, B, 2	M
	-2F	10,471	13	10	Slide torn	....
	-3F	10,492	17	11	II, B, 2	F
	-4F	10,558	9	9	II, B, 3	M
	-5F	10,617	604	20	II, B, 3 ?	MF
	-6F	10,647	20	14	II, B, 2	M
	-7F	10,649	9	13	II, B, 2	M
	-8F	10,652	25	13	II, B, 2	F
	-9F	10,656	45	12	II, B, 2	M
	-10F	10,805	24	16	II, B, 2	MF

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* Texture refers to the textural classification in table 1, p. 13.

† Average grain size refers to the approximate modal diameter of sand grains as estimated from thin-section examination. F = fine sand, 125–250 microns; M = medium sand, 250–500 microns; C = coarse sand, 500–1,000 microns; double letters (MF and MC) refer to sands on the border line between medium and fine or medium and coarse, respectively.

pores by secondary minerals; that is, a change in porosity from more than 25 per cent to between 10 and 15 per cent and a change in permeability from more than 2,000 millidarcys to approximately 10 millidarcys. During cementation by the process suggested, porosity would decrease progressively and slowly, not only because the pore space is progressively filled, but also because the solution at points of contact between grains permits closer packing. The process might proceed farther than is indicated by the present samples, perhaps reaching essential completion, or it might be halted at some intermediate point; but wherever it occurs, a very significant decrease in porosity and permeability is to be expected. In clay-cemented sandstones a decrease in porosity and permeability will certainly result from compaction. The magnitude of this reduction cannot be estimated from the present study.

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THE FACIES CLASSIFICATION OF ROCKS: A CLUE TO THE ORIGIN OF QUARTZO-FELDSPATHIC MASSIFS AND VEINS[1]

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ABSTRACT

The formation of quartzo-feldspathic rocks—principally pegmatites—is discussed from the viewpoint of thermodynamic stability of minerals and mineral associations.

Most quartzo-feldspathic rock bodies, migmatites, and pegmatitic veins encountered in folded chains exhibit mineral associations which indicate a formation at temperatures below which silicate melts cannot exist.

Data from a granulite facies area in western Greenland show that, even at the relatively high *P-T* conditions that prevailed there, the granitic component in the system was unable to form melts. Furthermore, pegmatites and quartzo-feldspathic massifs[2] usually possess mineral associations in harmony with the degree of metamorphism prevailing in the surrounding region. In high-grade areas the pegmatites—independent of their bulk chemical composition—show high-temperature mineral assemblages; in low-grade areas low-temperature assemblages occur in the pegmatites.

Because the melting curve of a magma is definitely determined by the bulk composition of the system (at constant pressure), the logical conclusion must be that most pegmatites, veined gneisses, migmatites, and several acid massifs were formed by some metamorphic-metasomatic processes in which silicate melts were unable to co-operate.

THE PROBLEMS

The origin of deep-seated quartzo-feldspathic rocks encountered in regional metamorphic areas is still an unsolved problem.

Very often the same quartzo-feldspathic rock or pegmatitic vein is interpreted as a magmatic crystallization product by one author, as a metamorphic-metasomatic product by another, while a third considers it to be the product of an anatectic pore melt or granitic ichor.

If there were a large overlap of the *P-T-X* (*X* = composition) fields of rock-making silicate melts and of solid minerals, the different viewpoints might be explained by supposing that the several authors defend somewhat one-sided views on a complex phenomenon in part magmatic and in part metamorphic. (The term "metamorphism" here comprises all endogenic processes taking place at submagmatic *P-T-X* conditions.) However, we have good experimental as well as field reasons for believing that magmatism is confined to temperatures above 600°–900° C. (depending on the content of water and other "volatiles" in the silicate system). Only above these temperatures should silicate minerals and melt coexist.

Below these temperatures the melt may be substituted for by saturated but dilute aqueous fluids. These saturated solutions should contain less than 1 or 2 per cent of silicates because common rock-forming minerals are not very soluble in water (Ingerson and Morey, 1937). On the other hand, silicate melts cannot absorb more than 10 per cent of water (Goranson, 1931, 1936, 1937); consequently, the system: rock-making silicate melts—water does not form a continuously miscible series.

The liquid or gaseous solutions occurring in pores and fissures in crystalline silicate rocks certainly may be charged with such easily soluble constituents as NaCl, KCl, alkali silicates, various solu-

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[1] Manuscript received March 19, 1948.

[2] The term "massifs" as used in this paper means large, homogeneous bodies of deep-seated rocks.

ble fluorides, sulphates, etc. (Newhouse, 1932). However, as such compounds practically never occur in the crystalline state in the common silicate rocks, we can conclude that the rock solutions are very seldom saturated in these constituents. Furthermore, it has never been proved experimentally or theoretically that aqueous solutions of such easily soluble constituents are completely miscible with the rock-making silicate melts. Therefore, in harmony with experimental data, we can draw a simple diagram (fig. 1) showing the *P-T* fields of magmatism and metamorphism, respectively.

In the magmatic field, melts and gases are involved in the formation of rocks. In the metamorphic field, the rock-forming processes take place in the solid state with or without a coexisting dilute fluid which is very different from a silicate magma both in composition and in physical properties.

It is reasonable to expect, therefore, that the rock or mineral assemblages formed from magma—either by magmatic differentiation or by partial anatexis and subsequent crystallization—would be quite different from those formed by metamorphism. Only if progressive metamorphism brings a rock to the boundary between metamorphism and magmatism will the products of these two processes be identical.

There are seldom different opinions as to whether basic rocks are metamorphic or magmatic. Regarding the metamorphic basic rocks, however, it is in many cases difficult to decide whether they were emplaced in a magmatic or in a submagmatic state. A given massive or foliated low-grade amphibolite or greenstone, which everybody agrees is metamorphic, may be interpreted as an ortho-amphibolite, as a para-amphibolite, or as a metasomatic amphibolite by different authors; but nobody will consider this low-grade rock as crystallized directly from a magma.

The case of quartzo-feldspathic rocks is quite different. Some investigators believe that these rocks have crystallized directly out of a magma; others believe that metamorphic recrystallization alone has taken place. The reason for disagreement is clear. The quartzo-feldspathic minerals are stable within a wide range of pressures and temperatures, from those of magmatic rocks down to the low greenschist facies. The common minerals in a granite or an acid quartz-diorite, therefore, are not indicative of the *P-T* conditions at which these rocks are formed. However, in several cases it is also possible to determine the mineral facies or pressure-temperature conditions which quartzo-feldspathic rocks represent (Eskola, 1921, 1939). These determinations suggest that most granites, granodiorites, and pegmatites encountered in gneissic areas formed at pressures and temperatures so low that very few of these rocks can be interpreted as primary magmatic products.

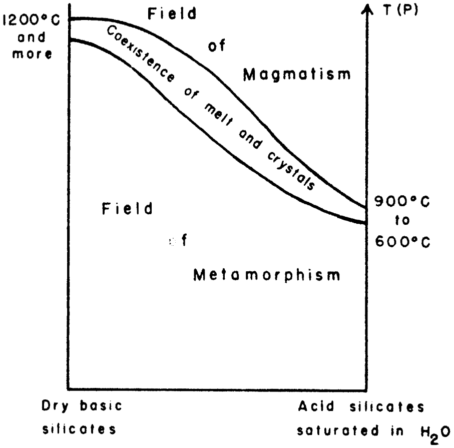

FIG. 1.—Schematic diagram of the relation between magmatism and metamorphism.

Some of the objections to the view that most quartzo-feldspathic bodies and veins are the product of a "progressive" magmatic differentiation from a juvenile basic magma support the "regressive" magmatic differentiation hypothesis. The partial anatexis and palagenesis of J. J. Sederholm, P. Eskola, Tom. F. W. Barth, and others explain the formation of some granitic or granodioritic bodies, some pegmatitic veins, and the phenomenon of granitization by assuming that during progressive metamorphism, a regressive magmatic differentiation takes place, that is, the easily melting components form a pore silicate melt which gradually absorbs the "basic" components in the rock complex. At a certain favorable stage this re-fused material is supposed to generate larger homogeneous bodies and pegmatitic veins. This hypothesis of regressive magmatic differentiation seems to take proper account of two of the main objections to juvenile magmatic differentiation: (1) the ratio between deep-seated acid and basic rocks is opposite to that between acid and basic lavas and (2) the space problem.

However, there are several objections to both these magmatic hypotheses. The problems fall into two main groups, namely, (1) the kinetic-structural problems and (2) the thermodynamic stability problems.

The kinetic-structural problems involve the movement or transfer of the rock-forming substances: Has it been a mechanical bodily motion or a diffusion? The thermodynamic stability problems involve the state of the system from, or in, which the problematic quartzo-feldspathic rocks form.

THE KINETIC-STRUCTURAL PROBLEMS

Let us consider only briefly the kinetic-structural problems. Coarse-grained and isotropic fabric of rocks is commonly considered as evidence of magmatic origin. Nevertheless isotropic coarse-grained massive rocks can form during metamorphic recrystallization or metasomatism, if the rocks are affected by confining homogeneous pressure and given a sufficiently long time of reaction. The epidote-bearing quartz-dioritic or granitic helsinkites and unakites, for example, have been interpreted as primary magmatic rocks (Mäkinen, 1916; Laitakari, 1918; von Eckermann, 1925; and Wilkman, 1928), but the mineral assemblages of these rocks positively indicate a *P-T* field well below that of hydrous silicate melts.

Likewise, numerous "magmatic" cross-cutting and conformable veins of isotropic fabric (some pegmatites, quartz veins, etc.) contain a mineral assemblage indicative of submagmatic conditions. It appears that both the coarse structure and the cross-cutting character of such veins have in many instances been the principal evidences on which investigators have based their conclusion of magmatic origin. For example, Frank L. Hess wrote in 1925 (p. 290), referring to the origin of pegmatites: "In attempts to explain their origin, all investigators seem to be agreed on one point, namely, that the pegmatites are of igneous origin." This was a reasonable statement before Goranson had made his experiments with solubility of water in granitic melts and before we knew much about other possible forces and processes which could form pegmatitic bodies, but Hess's viewpoint is no longer tenable. Even before the limited effect of water on the depression of the melting "point" of magmas was known, investigators (for example, Hess, 1925; Landes, 1933) observed evidences of metasomatic alteration of pegmatites. The commonly accepted conclusion was that the "simple" pegmatitic bodies underwent metasomatic alteration after their crystallization from a magma. This alteration was then assumed to be due to magmatic rest solutions of unknown composition and physicochemical character.

Commonly, larger bodies of quartzo-feldspathic rocks encountered in gneissic and regional metamorphic areas show foliation which is often explained as flow in a magmatic melt containing some early-formed crystals (Cloos, 1936). It is difficult to determine whether a foliated structure is due to flow in partly crystallized magma or in a solid rock. However, in a partly crystallized moving magma the parallelization of the minerals should be caused by form orientation of elongated or flaky minerals. On the other hand, deformation of the crystal units (e.g., quartz and feldspar) so commonly observed in deep-seated rocks implies a crystalline state of the rock body itself at the time of deformation.

If the foliated structure proves deformation in the solid state, the rock may, of course, be a recrystallized orthogneiss. The texture of deep-seated rocks, therefore, cannot be used as positive proof of the state of the substances during the emplacement of the rock material. Common features of folded mountain chains are quartzo-feldspathic massifs exhibiting structures conformable with their gneissic or schistose surroundings. Such conformable contacts show that the differences in fluidity among the several rocks have been small, as might be expected, if the folded and deformed complex were a complex of crystalline rocks rather than rocks intermingled with quartzo-feldspathic melts. If the general conformity is explained by assuming that the high pressure makes the viscosity of the melts approach the viscosity of the crystalline rocks, new difficulties for the magmatic hypothesis will arise. The magmatic hypothesis must assume that granitic melts—eventually charged with up to 10 per cent of water and other volatiles—are able to penetrate crystalline rocks during regional metamorphic conditions and form confined pegmatitic bodies in nongranitic rocks, e.g., amphibolites (Andersen, 1931). This permeability seems to be so great that it cannot possibly be due to any circulating pore ichor possessing a viscosity comparable to that of the crystalline country rock.

It seems impossible to me for any liquid (even water) to penetrate mechanically most metamorphosed rocks because they are compact and contain only an intergranular or mosaic pore system of molecular dimensions. Supermolecular fissures and ruptures which in places open up during mechanical deformation of rocks would be discontinuous and therefore unable to conduct circulating fluids. Furthermore, supermolecular openings in rocks are thermodynamically unstable and will be closed rapidly by plastic flow and diffusion in the strained zone around such fissures.

The larger fissures which extend throughout metamorphic terranes are, of course, able to convey magmas as well as watery solutions provided that the fissures are in communication with the proper sources. However, pegmatites are commonly not in communication with fissures of this type.

Lenticular conformable pegmatites so commonly encountered in gneissic schists (for example, the Black Hills pegmatites [Fisher, 1942, 1945]) appear to have pushed the country rock in front of their expanding boundaries, thus proving that the mechanical pressure in the pegmatite bodies has been greater than in the surrounding gneisses. Consequently, the introduced material has, during the formation of the pegmatite, moved from low toward high mechanical pressure. This would be physically impossible if the transferred material were a melt or hydrous liquid. Diffusion of individual atoms, molecules, or ions, however, can easily explain the formation of these pegmatites because molecular particles will automatically diffuse toward high mechanical pressure when the chemical potential, or the vapor tension, is lower at the site of high pressure than in the surroundings (Ramberg, 1944*b*, 1944*d*).

If silicate melts are able to penetrate crystalline rocks in a fashion required by the theory of magmatic origin of pegmatites, veined gneisses, etc., why is basic magma never found penetrating metamorphic rocks in the same diffuse manner? It is well known that basic magmas are more fluid than acid melts, and undoubtedly water and other volatiles decrease the viscosity of a basic magma just as they decrease the viscosity of acid silicate melts. It seems obvious that some kind of transfer of substances other than the simple bodily flow of magmas and solutions is required.

THE PROBLEMS OF THERMODYNAMIC STABILITY

Pegmatites as key rocks.—The mineral assemblages of some quartzo-feldspathic massifs belong to the same facies as that shown by the country rock. This may be explained by postmagmatic metamorphic recrystallization, although it is not clear why such rocks, if of magmatic origin, should be subjected in all cases to regional metamorphism.

However, pegmatites (commonly considered as crystallized rest liquors from granites) usually occur as "primary" vein bodies only rarely found deformed. It is therefore important to note that these "primary" pegmatitic rocks often contain a mineral suite that not only indicates submagmatic conditions but also denotes the same *P-T* values as those found in the wall rock. Thus, even if pegmatites are magmatic, considerable mobility in subsolidus silicate systems is required to explain the ability of coarse-grained, commonly cross-cutting pegmatites to recrystallize in a completely new metamorphic form.

It is generally assumed that pegmatites are due to watery rest solutions expelled from crystallizing magmas. In this connection, let us consider a comparatively low-grade metamorphic area where the "mother" granite is a recrystallized unakite with epidote, acid plagioclase, microcline, quartz, micas, etc., and the cross-cutting, undeformed pegmatites are of granodioritic composition. The metamorphic assemblage in the granite indicates a temperature perhaps some hundred degrees below the solidification "point" of any hydrous granitic magma (fig. 2). The hypothetical watery pegmatitic solution which crystallized as cross-cutting veins during the last stages of metamorphism must, if it was expelled from the granitic magma, have been wandering throughout the region from the time of crystallization of the magmatic granites to the consolidation of the pegmatites. Moreover, the unakites represent rather low degrees of metamorphism; and, therefore, the hypothetical pegmatitic solutions must have persisted during a long period of regressive metamorphism. Thus they must have interacted continuously with the metamorphosed minerals, and I see no way to decide now whether these solutions were of magmatic or metamorphic-metasomatic origin.

The importance of pegmatites in the problems under discussion makes a brief description of these rocks desirable. It is well known that there exists every transition among the several types of pegmatitic veins or segregations in regional metamorphic areas. In composition they range from basic oligoclase pegmatites through granodioritic, syenitic, and true granitic pegmatites to pure quartz bodies. Other more rare types are, on the one hand, the Li pegmatites and, on the other, the phlogopite-apatite-pyroxene-hornblende pegmatites.

Although the prevailing view is that many of the rare minerals and elements so commonly found in pegmatites are characteristic of the latest crystallization product of a magma, this view cannot be considered proved adequately. As the magmatic origin of pegmatites was not questioned (Hess, 1925), it was assumed that the occurrence of rare minerals and elements in them proved that the magmatic rest solutions carried these elements. Most rocks, sedimentary as well as igneous and metamorphic, contain several of the rare elements in spectroscopically detectable amounts (B, U, V, Be, etc., in sediments; Sc, etc., in amphibolites, etc.), so we usually have no difficulty in ascribing sources for most rare elements encountered in pegmatites. An enrichment of such rare elements in certain pegmatites by means of metamorphic-metasomatic processes (selective diffusion) is a priori no more impossible than progressive or regressive magmatic differentiation.

However, the idea of some leaching from the country rocks has been accepted only if the mineralogical composition of a pegmatite (garnet, cordierite, andalusite, kyanite, etc.) deviates too much from that reasonably expected of magmatic rest liquors (Andersen, 1931; Bjørlykke, 1934; etc.).

The form and the structure of pegmatites change within wide limits. Pegmatites in places penetrate gneisses in the form of plane-parallel dikes, but these are of limited extent as compared with true magmatic dikes such as rhomb porphyry, diabase, etc. Many pegmatites occur as confined lenticular bodies of either conformable or cross-cutting character. Completely irregular masses are also common. Pegmatites are most commonly sharply defined bodies, although gradational boundaries exist between the pegmatite and the country rocks.

The size ranges from microbodies to masses several hundred meters across. However, pegmatites larger than 5–10 m. wide and some hundred meters long are comparatively rare.

The texture within a pegmatitic body may be heterogeneous. A feldspar border phase and a quartz core are typical but are absent from many pegmatites.

The mineral assemblages denote a very wide *P-T* range from about 600° C. to perhaps 100° C., or from the granulite- or pyroxene-hornfels facies down to greenschist facies.

Although exhibiting such a diversity of properties, pegmatites have many characters in common, some of them not easy to define. The one character which most pegmatites have is occurrence as rather small bodies—usually of quartzo-feldspathic composition—with textures and composition distinguishing them from the country rock.

Another significant characteristic which justifies grouping of this type of rocks is their occurrence in regionally metamorphosed areas, apparently having some important relation to regional metamorphism. This fact was emphasized by O. Andersen (1931), although he considered the pegmatites to be chiefly of magmatic origin.

Our present knowledge of mobility of

substances by diffusion under regional metamorphic conditions gives a more plausible explanation of the striking connection in space and time between metamorphism and formation of pegmatites.

METAMORPHISM IN FOLDED MOUNTAINS

The association of granites, gneisses, and pegmatites in regional metamorphic areas indicates that the problem of the origin of all these types of rocks can be solved only by field and petrological studies in more or less eroded folded mountains. Folded mountain chains vary widely in character, but there are some important properties that all folded mountains have in common. On an average, for example, the degree of regional metamorphism increases from the low-grade superficial borders toward the high-grade central and deeper parts.

In the deep roots of eroded folded mountains the degree of metamorphism appears to be uniform over large areas. Thus, in the deeply eroded parts of the pre-Cambrian and Caledonian chains, rocks belonging to the epidote-amphibolite facies, the amphibolite facies, and the granulite facies cover thousands of square miles. A granulite facies area in western Greenland, for example, mapped by Danish expeditions in 1946, 1947, and 1948, extends about 150 km. across the strike between Holsteinsborg and Egedesminde. Investigations of the mineral associations show that the *P-T* conditions were astonishingly uniform throughout this vast area, although rocks vary widely in composition.

The vast anorthosite-birkremite field in western Norway described by C. F. Kolderup (1914, 1933) and Tom. F. W. Barth (1945) should also be interpreted as a regional metamorphic complex formed, or at least recrystallized, under *P-T* conditions corresponding to the granulite facies. Kolderup (1933) considered these rocks as "no doubt magmatic," although his description indicates that the whole field has been subjected to regional metamorphism, as there is no difference in mineral facies between the basic noritic rocks and the acid hypersthene granitic gneisses, massifs, and pegmatites.

A. W. Groves (1935) explains the hypersthene gneisses of Uganda, Africa, as a high-grade metamorphic facies of the surrounding "common" gneisses belonging to amphibolite facies. T. H. Holland (1900) interpreted the classic charnockite-norite complex of India as a vast magmatic series. His evidences were cross-cutting dikes and structural properties which now cannot be accepted as proof of magmatic origin. The norite dike (Holland, 1900, pp. 228–230) which cuts a gneiss lying in contact with the charnockite series is probably a diabase dike which was recrystallized, together with the host gneisses, in the granulite facies. There are several examples of the same phenomenon in western Greenland.

Holland's detailed rock descriptions show that—as in Uganda, Norway, Ellesmerland (Bugge, 1910), and western Greenland (tables 1, 2, 3)—the members of the norite-charnockite "kindred" correspond to the same *P-T* conditions. I am thus convinced that the classic Indian area contains a series of high metamorphic rocks belonging to the granulite facies.

The areal extent of the amphibolite facies and the epidote-amphibolite facies in eroded folded chains depends on the definitions of these facies. Pentti Eskola, who introduced the concept of mineral facies (1921, 1939), distinguished between the epidote-amphibolite facies and the amphibolite facies on a less accurate conception of the stability relationship between epidote and plagioclase (Ramberg, 1944*a*, 1944*b*). Epidote is stable at relatively higher temperatures than was supposed when the epidote-amphibolite facies was defined. More important, however, is the fact that intermediate and acid plagioclase are stable far below the uppermost limit of the stability field of epidote (fig. 2). Thus, if the epidote-amphibolite facies is to include the whole stability field of epidote, the epidote-amphibolite facies would have a very large areal extent in gneissic regions. The amphibolite facies, on the other hand, will be narrowly restricted because the *P-T* gap between the stability fields of rhombic pyroxene and epidote is rather small (figs. 2–3). It may be better, therefore, to set the boundary between the epidote-amphibolite facies and the amphibolite facies at the lower *P-T* boundary line of diopsidic pyroxene. This line corresponds approximately to an equilibrium between epidote and plagioclase of composition An_{30} under regional metamorphic conditions (fig. 2). Thus defined, the highest part of epidote-amphibolite facies (epidote in equilibrium with An_{20-30}) and the amphibolite facies cover very large areas in eroded folded mountains of pre-Cambrian, Caledonian, and Hercynian age.

In less eroded folded chains the change of degree of metamorphism is more rapid and irregular to the extent that we are above or outside the main field of quartzo-feldspathic gneisses. The degree of metamorphism in less eroded mountains is considerably more dependent on the large-scale tectonics so that, for example, highly metamorphosed complexes of gneisses are interfolded in, and thrust over, schists of a lower degree of metamorphism.

Pentti Eskola (1939) considered the greenschist facies, epidote-amphibolite facies, and the amphibolite facies as the "normal" facies of regional metamorphism, placing, among others, the granulite facies outside this "normal" sequence. However, investigations of several areas—Groves (1935), Kolderup's and Barth's descriptions of the Egersund area in Norway (1914, 1933, 1945), Adams' description of Ceylon (1930), Hietanen's description from Finland (1947), and my unpublished investigations in Greenland—show that the granulite facies should be placed in the "normal" sequence of regional metamorphism. This facies represents the highest *P-T* conditions that prevailed during regional metamorphism.

The granulite facies has, as we know, its greatest extension in old pre-Cambrian areas, although it is also found in younger chains. In the Caledonian chain in Norway, for example, the granulite facies is found in two situations: (1) in the roots of the chain in the west coast district (the Bergen areas and farther north) and (2) in the form of overthrust rocks: the Bergen-Jotun "kindred" of V. M. Goldschmidt (1916).

Even before the last-mentioned complex could be interpreted as metamorphic-metasomatic rocks formed in the granulite facies, Goldschmidt and others found that they were thrust—in the solid state—over less metamorphosed rocks. Their origin now seems clearer. They were formed, or at least recrystallized, under granulite facies conditions in the Caledonian root, along the west coast of Norway, and have not recrystallized to lower facies during their comparatively rapid tectonic elevation.

With increasing degree of metamorphism in folded mountains, the process of granitization (or perhaps, more correctly, quartzo-feldspathization) is intensified. Quartzo-feldspathic massifs, either in the form of large conformable batholiths (basement granites, diapire granites) or in the form of rather homogeneous folded layers among the more heterogeneous schists and gneisses, appear as one crosses the eroded mountains toward the central parts.

Very typical features are the pegmatitic bodies which apparently develop in the epidote-amphibolite facies (usually not far below the stability temperature of diopside) and continue throughout the epidote-amphibolite, the amphibolite, and the granulite facies.

Field work in western Greenland indicates that the pegmatites are typical of epidote-amphibolite and amphibolite facies areas, whereas in the granulite facies area the number and size of pegmatites seem somewhat less. For the Greenland area this can be interpreted as a pressure effect. Fissures and other low-pressure places activating pegmatitic crystallization do not easily form during the heavy pressure prevailing in granulite facies areas. However, whether or not this is a general phenomenon cannot be determined because no granulite facies area has yet been adequately described.

In greenschist facies areas and in the actinolite-schist facies the pegmatites seem to be represented by quartz veins and alpine veins which exhibit minerals of low-temperature origin. It is, however, worth while to note that in alpine veins, also, quartz and feldspars are the principal minerals, with some chlorite, calcite, etc. (Koenigsberger, Niggli, Parker, 1940).

The phenomenon of granitization and other types of metasomatism is highly varied and polyphase in detail. Whereas some basic and nongranitic rocks have been gradually imbued with quartzo-feldspathic material reaching, successively, the composition of a granite, a granodiorite, or a quartz-diorite, some marbles and amphibolites remain for a long while unaltered. In other cases rocks have even been basified by removal of quartzo-feldspathic material or by introduction of ferromagnesian elements. D. Reynolds (1946) has very clearly described and emphasized this process of basification. Quantitatively, basification plays a less important role than quartzo-feldspathization. Qualitatively, the process is an important detail of regional metamorphism and metasomatism.

The distances over which substances migrate in one manner or another through crystalline rocks, altering them here and there, and consolidating as confined pegmatitic veins at other places, are not easy to determine. We observe the result of an intricate complex of processes usually without any means of determining the earlier stages. As a rule we cannot definitely decide whether a quartzo-feldspathic gneiss is the result of mere recrystallization or of metasomatic replacement.

One of the first proofs of large-scale introduction of quartzo-feldspathic substances into schists was offered by V. M. Goldschmidt in his paper on the Stavanger area (1920). In addition to a gradation in rock composition at different distances from a granitic body, another proof of the penetration of crystalline rocks by foreign materials is the presence of pegmatites in metamorphic amphibolites and other nongranitic rocks. If structure alone be considered, migmatites and agmatites could be explained by the hypothesis of an invading or anatectic magma having intermingled with the old rocks. However, the usual mineral facies of migmatites, agmatites, and similar rocks belongs to the field of metamorphism and thus proves that the movement of matter has not taken place in the form of silicate melts.

THE MAGMATIC THEORY VERSUS THE METAMORPHIC-METASOMATIC THEORY

The foregoing brief description outlines the principal features of metamorphic areas. The main problem to be discussed is: Are granites and granodiorites mainly magmatic, i.e., intruded from juvenile or anatectic melts during the orogeny; and are the pegmatites the crystallized residual melts or anatectic liquor; and the migmatites, veined gneisses, etc., a solidified mixture of these melts and the remnants of the older rocks? Or are these features typical of most metamorphic areas, due principally to different processes—taking place largely in the submagmatic state?

The latter possibility may be briefly outlined as follows: Sediments and lavas overlying a "granitic" substratum in a geosyncline are folded and deformed together with the substratum itself. The "younger" granites and granodiorites in the folded chain are then partly granitic substratum, which is interfolded and intruded in a crystalline state into the superlayers, and partly crystallized arkoses and acid lavas. The acid rocks are in part formed by metasomatism, which involves diffusion among the different chemically incompatible rocks, and up from the granitic substratum. According to this theory, the pegmatites are formed by metamorphic-metasomatic processes without the participation of silicate melts.

This theory, of course, does not deny the occurrence of granitic or acid melts where the temperature was high enough. But, because the temperature of even the highest grade of metamorphism (granulite facies) is less than that of magmatism, the theory considers magmatism negligible in the formation of acid rocks found in regionally metamorphosed areas.

Several reasons for this view will be discussed later. Field observations strongly impress one with the continuity between rather homogeneous granitic massifs and low-grade schists. Such a continuity would not be the case if the rock-forming agency were in part a concentrated granitic melt and in part a very dilute watery solution. However, if the processes were all submagmatic, a transition from unaltered to completely granitized rocks would be expected.

The common regional increase of temperature is supposed to be due in part to the geothermal gradient and to radioactivity, in part to exothermic chemical reactions between incompatible rocks, and in part to the liberation of heat of consolidation where the activated diffusing atoms, molecules, and ions consolidate.

The requirements of the magmatic hypothesis should be that (1) the granodioritic-granitic substances are introduced in the form of silicate melts; (2) there is in every case evidence that mechanical forces have acted on the melt to move it in the required direction; (3) the melt can penetrate solid crystalline rocks to form veined gneisses, "augen"-gneisses, pegmatites, and granitized rocks; (4) the mineral assemblages of the granitic rocks and veins correspond to the solidus temperature of the magma in question and are the same for all rocks of the same composition (with the same content of "volatiles") and vary with composition according to the rules of magmatic crystallization; and (5) the temperature in the crystallizing massifs and pegmatites as a rule be different from that in the regional-metamorphic surroundings (because the crystallization temperature of a magma is determined by its composition and pressure and must be independent of the physical and chemical conditions in the surroundings).

The requirements of the metamorphic-metasomatic hypothesis are that (1) the quartzo-feldspathic bodies be formed below the field of magmatism, i.e., below approximately 550°–900° C.; (2) the mineral association in the "granitic" bodies (massifs and pegmatites) correspond approximately to the degree of regional metamorphism in the area where such bodies are situated (some smaller or larger deviations are to be expected owing to chemical instabilities, to exothermic and endothermic interactions, and to different velocities of reactions of different rocks and minerals); (3) the temperatures of formation of the rocks (indicated by the mineral assemblages) be independent of the chemical composition; (4) there be some relation between chemical composition of a pegmatite and the enclosing rocks; (5) in every case there exist chemical potential differences which can explain the diffusion of atoms, ions, and molecules through rocks in the required directions (if supermolecular fissures exist, flow of watery solutions or flow of gases will take place); (6) the diffusion coefficients for the several atoms, ions, or molecular compounds in rocks be considerably different, so that selective diffusion can explain metamorphic-metasomatic differentiation (for example, the formation of granite pegmatites in amphibolites); (7) the diffusion through the different pathways in rocks be rapid enough to produce the observed metasomatism; and (8) the crystalline plasticity be sufficient to permit the considerable movement which takes place during formation of most migmatitic gneisses. The fluidity of the quartzo-feldspathic material must be greater than that of the basic rocks.

Points 1, 2, 3, and, in part, 4, are discussed here, but points 5–8 are expounded more thoroughly in other papers.

The solution of the problem lies in the determination of the *P-T* conditions of formation of various rocks and mineral assemblages. The most adequate method seems to be the investigation of stable mineral associations according to the principle of mineral facies, although every type of geological thermometer may be used.

In most deep-seated regional metamorphic and plutonic rocks the minerals are to a great extent in mutual equilibrium within more or less confined parts of the rocks. The mineral facies is governed by the *P-T* conditions extant during the main period of formation. Minerals formed at *P-T* conditions different from those of the main assemblage are usually not difficult to recognize.

These observations on deep-seated rocks are not always easy to interpret. Special difficulties are presented by the granulite facies formed in the deeper and/or central parts of folded mountains, which retain their high mineral assemblages during elevation of the crust and are often overthrust on low-grade rocks. Metamorphism is by no means a continuous and reversible process which is progressive when the rocks (sediments, lavas, etc.) are depressed and folded down in the "crust" and regressive when they are elevated. Instead, during a period of orogeny, conditions permit metamorphic-metasomatic reactions which reach chemical equilibrium in certain parts of the crust. After these conditions disappear, the power of reaction decreases rapidly enough to prevent extended retrograde metamorphism.

Absence of retrograde metamorphism may be explained in three ways. (1) Hot magmas may be injected, keeping the temperature in the surroundings high and rather constant for a long time. After the melt has crystallized, there follows a decrease in temperature too rapid to permit mineral readjustment to the new conditions. (2) Catalysts may promote the chemical reactions during the course of the metamorphism. These catalysts may subsequently lose their effect or disappear. (3) Heat liberated during the mechanical and chemical readjustments accompanying the formation of folded mountains may be the main factor in regional metamorphism. In the later stages of the mountain-making processes (the cratogenesis) these readjustments cease and no more heat is liberated, so that the power of reaction falls rapidly with the decreasing temperature.

The temperature declines because of (1) cessation of the exothermic chemical and mechanical equilibriopetal processes and (2) rapid elevation through colder zones extinguishing chemical action.

The power of metamorphic interaction, K, can be expressed as follows:

$$K = A \cdot e^{-q/RT}, \qquad (1)$$

where R is the gas constant, T absolute temperature, A a constant depending on the type of interaction, and q is the molal heat of activation. The quantity q is usually very high, of the order of magnitude 20,000–50,000 cal. (Barrer, 1941). Decreasing temperature, consequently, greatly retards the velocity of interaction between solids.

Calcoferromagnesian minerals undergo many transformations within the wide P-T field between greenschist facies and granulite or pyroxene-hornfels facies. At the same time, quartzo-feldspathic rocks remain seemingly unaltered because quartz, potash feldspar, and acid plagioclase are stable minerals within the whole P-T range mentioned. Biotite and muscovite, the other two important minerals in quartzo-feldspathic rocks, are also stable throughout most of the P-T field of regional metamorphism. Biotite is stable at favorable chemical conditions in parts of the greenschist facies as well as in granulite facies. Biotite is also stable in the pyroxene-hornfels facies, and undoubtedly this mineral can crystallize directly out of a magma. However, the chemical stability field of biotite seems to be small in both the greenschist facies and in the granulite facies.[3]

In the epidote-amphibolite facies and in the amphibolite facies the P-T conditions are most favorable for formation of biotite, and the chemical stability field of biotite is extensive.

The narrowing of the chemical stability field of biotite in the granulite facies is due to the partial splitting of some end members of the biotite series into hypersthene and/or garnet with the liberation of potash feldspar, water, etc.

Muscovite is stable from the lowermost temperatures of metamorphism up to the boundary between the amphibolite facies and the granulite facies, where the mineral is changed to sillimanite and potash feldspar.

The third calcoferromagnesian mineral which occurs in some quartzo-feldspathic rocks is hornblende. Under favorable chemical conditions this mineral is stable throughout the whole range from the actinolite-greenschist facies up to the granulite facies, where it gradually splits up into hypersthene, plagioclase, etc., so that only certain types of hornblende can exist up to the field of magmatism.

The chemical stability field of hornblende, however, changes markedly throughout the different mineral facies. The field is narrow in the lowest grades of regional metamorphism and in the granulite facies, but in most of the epidote-amphibolite facies and in the whole of the amphibolite facies hornblende has a very large chemical stability field.

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[3] The chemical stability field of a mineral is definable as the field, or polydimensional space, within which the bulk composition of an assemblage can vary without disappearance of the mineral in question.

Between the amphibolite facies and the granulite facies only aluminous hornblendes rich in iron and/or titanium are stable in quartzo-feldspathic rocks (p. 36). In unsaturated rocks even hornblendes poor in TiO_2 and FeO or Fe_2O_3 are stable in the granulite facies.

The most obvious indication of the change from the amphibolite facies to the granulite facies, or vice versa, in quartzo-feldspathic rocks is no doubt the appearance or disappearance of orthorhombic pyroxene.

Although potash feldspar and acid plagioclase are stable from diagenetic conditions up to magmatism, the feldspars undergo distinct changes in passing from the amphibolite facies into the granulite facies. In granulite facies typical perthite and antiperthite are common (see description by Holland, 1900; Goldschmidt, 1916; Groves, 1935; Adams, 1930; Gevers and Dunne, 1942).

The plagioclase feldspar in the granulite facies has a characteristic greenish-brown color.[4]

In conclusion, there are distinct indicators of the *P-T* change from the amphibolite facies to the granulite facies in most quartzo-feldspathic rocks, but rocks of restricted composition (for example, some biotite granites) do not exhibit any distinct change in mineralogy when they are changed from the amphibolite facies into the granulite facies, or vice versa.

THE EPIDOTE-PLAGIOCLASE AND THE BIOTITE-GARNET EQUILIBRIA

In the large area from the greenschist facies to the granulite facies the most distinct *P-T* indices in quartzo-feldspathic rocks seem to be (1) the continuous equilibrium between plagioclase and epidote or zoisite and (2) the continuous equilibrium between Mn, (Fe, Mg)-garnet plus potash feldspar, and biotite (plus muscovite or sillimanite).

These two continuous equilibrium systems register the whole *P-T* range from the greenschist facies throughout the epidote-amphibolite facies, and the amphibolite facies up to the granulite facies (Ramberg, 1944*a*).

To avoid any misunderstanding, I emphasize that at present only the principal features of these equilibrium systems are known. These known features are, however, sufficient for the use to which they are put in this paper. I hope that enough field and laboratory data can be obtained to give accurately the structure of these two systems.

The following two equations show the transition from anorthite to epidote in potash-rich (3) and potash-free (2) rocks, respectively:

$$\begin{array}{lllll} & \text{high temperature} & & \text{low temperature} & \\ & 2CaAl_2Si_2O_8 + \tfrac{1}{2}H_2O & \rightleftharpoons \tfrac{1}{2}Al_2SiO_5 & + Ca_2Al_3Si_3O_{12}(OH) & + \tfrac{1}{2}SiO_2, \\ & \text{anorthite} & \text{kyanite} & \text{epidote} & \text{quartz} \\ \text{density:} & 2.76 & 3.50 & 3.25 & 2.65 \\ \text{volume:} & 2\cdot 101 \quad \Delta V \sim 29\text{cc.} & \tfrac{1}{2}\cdot 50 & 139 & \tfrac{1}{2}\cdot 23 \end{array} \qquad (2)$$

$$\underset{\text{anorthite}}{4CaAl_2Si_2O_8} + \underset{\text{orthoclase}}{KAlSi_3O_8} + 2H_2O \rightleftharpoons \underset{\text{muscovite}}{KAl_3Si_3O_{10}(OH)_2} + \underset{\text{epidote}}{2Ca_2Al_3Si_3O_{12}(OH)} + \underset{\text{quartz}}{2SiO_2} \qquad (3)$$

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[4] Unpublished chemical tests indicate that the color is due to absorption of about 0.5 per cent FeO and Fe_2O_3 in the feldspar lattice. As the typical greenish-brown Fe-bearing feldspar in the granulite facies is always associated with Fe-rich silicate, such as hypersthene and some hornblendes, probably the high *P-T* conditions of the granulite facies enable iron to enter the feldspar lattice provided that the chemical potential or vapor tension of Fe in the rocks is high enough.

The transition temperature of equation (3) is expected to be somewhat higher than that of equation (2) because muscovite has, in association with potash feldspar, somewhat less partial Al-tension or activity than kyanite.

Figure *2A*, shows approximately the continuous equilibrium between epidote and plagioclase resulting from equations (2) and (3) when albite is taken into consideration. We may postulate that the curves begin at about 400°–500° C. in the Ca side of the diagram and end at room temperature in the Na side, and field experience indicates that the lines curve much as in figure 2.

It is important to note that in potash feldspar- and muscovite-bearing rocks the plagioclase which is in equilibrium with epidote should be somewhat more acid than in muscovite-free rocks at the same *P-T* conditions, the difference becoming negligible at lower temperatures. Figure 2 does not hold for calcite-bearing rocks.

Qualitatively, we can conclude that increasing H_2O tension in the rocks will displace the equilibrium toward epidote. The ability of epidote to take much Fe into its lattice makes increasing Fe/Al ratio in rocks displace the equilibrium somewhat toward the epidote side. Equations (2) and (3) show also that the equilibria depend on the SiO_2 content, so that in quartz-free rocks the epidote may de stable at higher temperatures than in quartzose rocks.

The only experiments which indicate something about the absolute *P-T* conditions of the epidote-plagioclase equilibrium is the hydrothermal synthesis of anorthite at comparatively low pressures (Niggli and Schlaepfer, 1914).

At 470° C. anorthite repeatedly was formed in these experiments at hydrothermal conditions, thus indicating that epidote or zoisite is not stable at, or above, 470° C. at comparatively low pressure.

The effect of pressure upon equations (2) and (3) cannot be computed correctly until the heat of reaction is known. Some rough estimates may be obtained by using reasonable values of the heat of dehydration of silicates.

Epidote is perhaps the silicate mineral which binds its water most strongly because this mineral, unlike most hydrous minerals, does not lose all its water below 1,000° C. (Smethurst, 1935). Hydrous minerals which apparently bind their water content with much less energy than epidote exhibit molal energies of dehydration of the order of magnitude 20,000–35,000 cal. (Eitel, 1941).

Thus, using a value of 24,210 cal. (equal to 1,000 liter-atmospheres) for the heat of transition of equation (2), we obtain from the Clausius-Clapeyron law:

$$\frac{dT}{dP} = \frac{T_u \Delta V}{Q} = \frac{743 \cdot 0.029}{1,000} = 0.021°/\text{atm.}$$

At a depth of 20 km., for example, the transition temperature should increase from 470° C. to about 600° C.

However, 470° is probably too high for the uppermost stability temperature of epidote at low pressure, and it is also likely that 24,210 cal. is too low for the heat of reaction of equation (2). Therefore, the temperature values given here might be above the highest stability boundary of epidote. At least, we have at the present no reason to believe that epidote is stable at higher temperature than those evaluated here.

The volume of water is disregarded in equation (2) because there is every reason to believe that water molecules can migrate rather easily through intergranular spaces in metamorphic rocks, so that the rock pressure does not equal the water vapor pressure.

The simplest equation for the reaction of potash feldspar and garnet during regional metamorphism is

$$\underset{\text{potash feldspar}}{KAlSi_3O_8} + \underset{\text{garnet}}{(Mg, Fe)_3Al_2Si_3O_{12}} + H_2O \rightleftharpoons \underset{\text{biotite}}{K(Mg, Fe)_3AlSi_3O_{10}(OH)_2} + \underset{\text{sillimanite}}{Al_2SiO_5} + \underset{\text{quartz}}{2SiO_2} \quad (4)$$

In this equilibrium the Mg/Fe ratio in garnet is purposely less than in biotite (see tables 5, 6), so that the Fe end members react according to equation (4) at somewhat lower temperature than do the Mg members.

Below the degree of metamorphism at which sillimanite reacts with potash feldspar with formation of muscovite, the reaction (4) is changed to

$$\underset{\text{potash feldspar}}{2KAlSi_3O_8} + \underset{\text{garnet}}{(Mg, Fe)_3Al_2Si_3O_{12}} + 2H_2O \rightleftharpoons \underset{\text{biotite}}{K(Mg, Fe)_3AlSi_3O_{10}(OH)_2} + \underset{\text{muscovite}}{KAl_3Si_3O_{10}(OH)_2} + \underset{\text{quartz}}{3SiO_2} \quad (5)$$

Now, taking the spessartite component of garnet into consideration, we arrive at an equilibrium diagram between (Mg, Fe, Mn)-garnet, potash feldspar, and biotite as shown in figure 2, *B*. In this diagram we know that the equilibrium curve will be displaced upward somewhat by an increase of the Mg/Fe ratio and depressed slightly by a decrease of the Mg/Fe ratio. The correct form for the equilibrium curve is not fully known; it is certain only that it begins in the granulite facies and ends down in the lowermost part of the epidote-amphibolite facies, or in the greenschist facies. (Rocks containing the minerals plagioclase, epidote, garnet, biotite, potash feldspar ± [muscovite and/or kyanite] would be excellent for determining the relative position of the curves in figure 2, *A* and *B*.)

FIELD AND PETROLOGIC OBSERVATIONS

The granulite facies is comprised of rocks which have been affected by the highest *P-T* conditions of regional metamorphism and granitization. This facies of rocks is therefore decisive; for, if we prove that the easiest melted rock-making silicate system (i.e., a granitic system of certain composition) has not been liquid in the granulite facies, it is obvious that silicate magmas cannot exist at lower temperatures of regional metamorphism. Consequently, we may then assume that granites, granodiorites, migmatites, veined gneisses, and pegmatitic bodies exhibiting minerals belonging to the amphibolite facies, and to the epidote-amphibolite facies are either thoroughly recrystallized or are formed by some metamorphic metasomatic processes.

It might perhaps be assumed that volatiles depress the melting "point" of quartzo-feldspathic systems down to temperatures characteristic of the amphibolite and the epidote-amphibolite facies even if we can prove that the rocks in question have not been liquid in the granulite facies. The granulite facies has been considered an exceptionally "dry" one, so that it would be natural to assume that the water content has been insufficient to make a granitic material melt (Goldschmidt, 1916). However, there are some phenomena which are inconsistent with this assumption. For example, under the magmatic theory the coarse structure of pegmatites and the apparent mobility of the pegmatitic material is due to a great concentration of water and other volatiles in the pegmatitic magma. However, in the Greenland pegmatites in

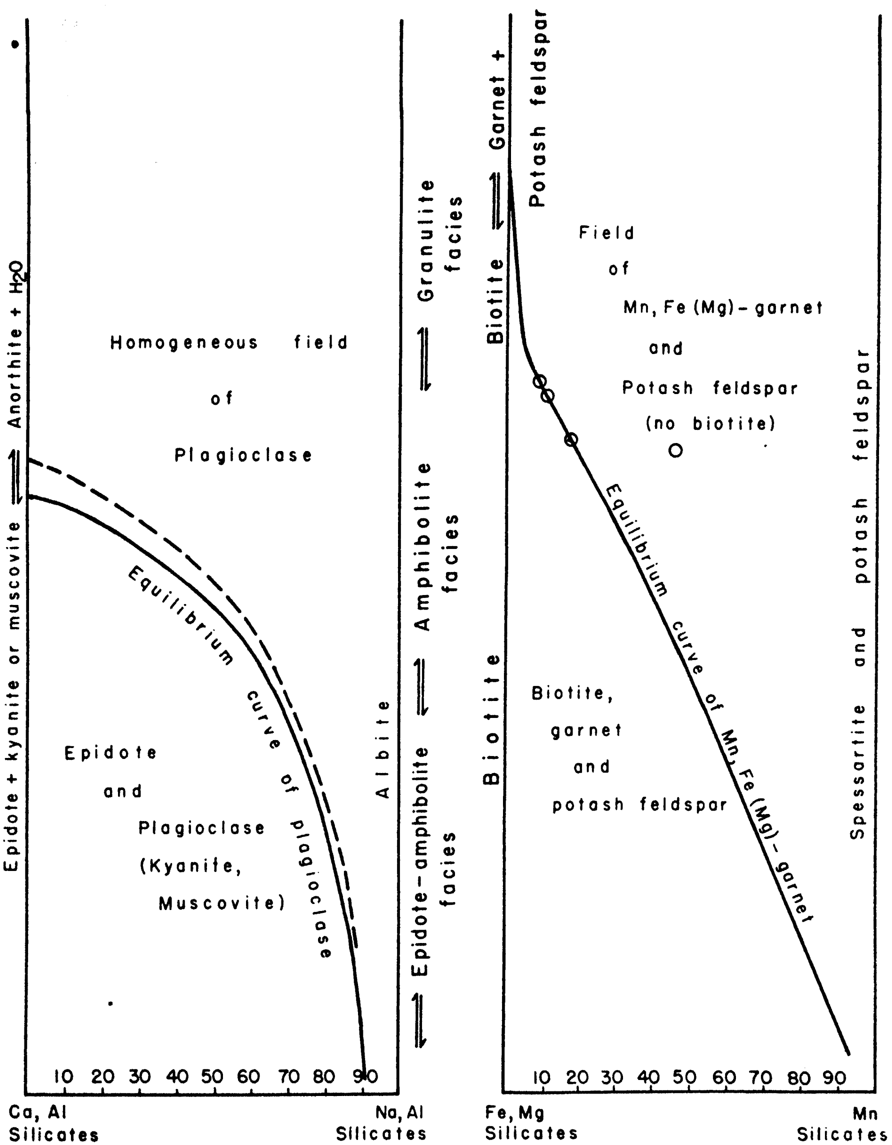

FIG. 2.—*A*, Subsolidus diagram of the plagioclase ⇄ epidote equilibrium. The composition of the system which the diagram represents corresponds to equations (2) and (3), p. 30, with the exception that Ca can be substituted for by Na. The transition temperature between anorthite and epidote corresponds to about 500° C. The solid curve gives approximately the composition of plagioclase in equilibrium with epidote in K-free rocks, whereas the stippled curve represents plagioclase in equilibrium with epidote in rocks containing muscovite and/or potash feldspar.

B, Subsolidus diagram of the biotite ⇄ potash feldspar, garnet equilibrium (eqs. [4] and [5], p. 32). The weight percentage of spessarite in garnets in equilibrium with biotite, potash feldspar (sillimanite, muscovite) is indicated by the equilibrium curve. Garnets from pegmatites in amphibolite facies gneisses in western Greenland are plotted in the diagram (see table 6).

the granulite facies area the pegmatitic mineral suite also corresponds to the granulite facies, proving that the crystallization temperature of a hypothetical hydrous pegmatitic magma was at least not below the temperature prevailing in the granulite facies.

The assumption that granulite facies conditions are not favorable for the introduction of volatiles and water into an anatectic melt seems unreasonable. The absorption of water in a silicate melt increases with water vapor pressure, so that, at places where the water vapor pressure is high, conditions should be most favorable for the formation of silicate melts by increasing temperature.

It seems to me that we must assume that the water vapor pressure in the pore system in rocks is usually higher in the granulite facies than in lower facies despite the smaller content of water in the granulite facies minerals. The expulsion of water from chlorite, biotite, anthophyllite, etc., when they are altered to garnet and hypersthene is due to an increase of the water vapor pressure of hydrous minerals with increasing metamorphism. Thus, in a garnet-hypersthene rock formed in the granulite facies from a chloritic, talc, anthophyllite-bearing low-grade rock, the water vapor pressure in the pores will be higher than in the water-rich low-grade rock. Because both temperature and water vapor pressure are higher in the granulite facies than in lower facies, the granulite facies should actually represent the most favorable conditions for formation of anatectic pore melts.

There is no reason to believe that juvenile magmas are poorer in volatiles where they are intruded into a granulite facies area than where they are intruded in rocks belonging to lower facies. In fact, the opposite is more probable because the high pressure responsible for the granulite facies should prevent more completely the escape of volatiles. Consequently, there seems to be no reason to believe that the melting temperature of a rock-making silicate system is lower in the amphibolite and the epidote-amphibolite facies than in the granulite facies.

Adequate data are not available concerning absolute pressures and temperatures that give rise to the facies.

Wollastonite which has been found in the granulite facies (Adams, 1930; Hietanen, 1947) and also in the highest parts of the amphibolite facies (Eskola, 1915; von Eckermann, 1922) cannot be considered an adequate geological thermometer. V. M. Goldschmidt's well-known calculations give only the highest temperature limit for wollastonite at different pressures (Ramberg, 1946*b*). Wollastonite can form at temperatures below 500° C. regardless of the rock pressure because CO_2 is able to migrate through the pores in metamorphic rocks so that the CO_2 pressure is never fixed in metamorphic rocks. The fact that wollastonite as well as calcite and quartz occur together both in the granulite facies and in the highest parts of the amphibolite facies is in complete harmony with the theoretical considerations outlined above. (R. Kühne, 1932, records wollastonites in association with epidote in Auerbacher marbles.)

The only data on the temperature of formation of the granulite facies are H. von Eckermann's determination for the highest part of the amphibolite facies in Mansjö in Sweden (1922). The association of α- and β-quartz shows the temperature of formation to be about 600° C. This value may serve as the lower boundary of the granulite facies. According to this rather uncertain estimate, hydrous granitic melts might or might not be stable in the granulite facies of rocks.

However, we shall turn to field, microscopical, and chemical studies of the Greenland rocks and minerals in deciding whether or not granites and pegmatites have been melted in the granulite facies.

GRANULITE FACIES AREAS

A. W. Groves (1935) seems to have been the first to prove that a series of rocks of charnockitic affinity, in Uganda, Africa, are the granulite facies metamorphic derivatives of gneisses belonging to amphibolite and epidote-amphibolite facies. My work on the pre-Cambrian of western Greenland shows that the hypersthene gneisses (enderbitic gneisses) and the noritic rocks there must be interpreted in the same way as the hypersthene gneisses in Uganda.

Rocks belonging to the granulite facies strike W.SW. and cover an area of about 15,000 sq. km. west of the ice cap between Holsteinsborg and the Arfersiorfik fiord south of Egedesminde. North and south of this area are rocks of the amphibolite facies and gneisses belonging to the epidote-amphibolite facies. The rocks of the granulite facies are mostly rather homogeneous enderbitic gneisses containing a certain bluish quartz, brownish-green andesinic antiperthite, orthorhombic pyroxene, ilmenite, and magnetite as the principal constituents. Perthitic potash feldspar, diopsidic pyroxene, garnet, a peculiar Fe- and Ti-rich hornblende, and biotite are also commonly found in the enderbitic gneiss. Only rarely is the potash feldspar content sufficient to make the gneisses mangeritic (Kolderup, 1933) and charnockitic. In this gneissic complex are found bands, lenses, and more equidimensional bodies of basic rocks usually of hornblende noritic composition. Folded layers of marbles, khondalites, kinzigites, and graphite gneisses occur conformably alternating with the acid hypersthene gneiss. Large and small conformable and cross-cutting pegmatites of heterogeneous mineralogical and chemical compositions are commonly scattered in the several gneisses throughout the whole area.

The structure of the complex and apparently also the chemical composition of the rocks are the same as are commonly observed in gneissic areas belonging to the amphibolite and the epidote-amphibolite facies.

Of importance are several noritic bands and lenses in the acid hypersthene gneisses which are interpreted as the granulite facies derivatives of the amphibolitic and epidote-amphibolitic layers, boudins, and deformed dikes found in the gneisses both south and north of the granulite facies area. Farther south along the coast the amphibolitic layers are in turn found to be cross-cutting diabase dikes.

The noritic components in the hypersthene gneiss are, therefore, derived from common diabases. The hypersthene-bearing enderbitic gneiss that commonly includes and penetrates the noritic layers is an activated granulite facies variety of the old hornblende-biotite gneiss which once was invaded by the diabase dikes.

These field observations show that the rocks of charnockitic affinity (norites, enderbites, charnockites) were not formed by magmatic differentiation but by high-grade regional metamorphism. In addition to these briefly stated field evidences, we shall consider the mineralogical evidences in detail.

In the norite-enderbite-charnockite series of Greenland as well as in rocks of other areas (Washington, 1916; Bugge, 1910; Groves, 1935; Parras, 1941; Gevers and Dunne, 1942) the basic members are rich in hornblende whereas orthorhombic pyroxene is the most common ferromagnesian mineral in the acid members. (Diopsidic pyroxene, biotite, and garnet are commonly found in basic as well as in acid members.)

As emphasized by Groves (1935, pp. 194–200), this is not in accord with the theory of magmatic differentiation, according to which hornblende should crystallize later and at lower temperatures than should hypersthene. According to the facies diagrams of Eskola (1939), we should also suspect our basic hornblende noritic rocks of having consolidated at a lower temperature than the surrounding hypersthene gneisses. The explanation is that all types of rocks are recrystallized, and in part formed, at very nearly the same *P-T* conditions. It is the stability relationships of hornblende which make this mineral form in basic rocks at the same temperature at which hypersthene forms in acid rocks.

The Al-free actinolitic and tremolitic hornblendes are unstable in the highest parts of the amphibolite facies and in the granulite facies; they split up into hypersthene and diopsidic pyroxene:

$$\overset{\text{low temperature}}{\underset{\text{actinolite}}{Ca_2(Mg, Fe)_5Si_8O_{22}(OH)_2}} \rightleftharpoons \overset{\text{high temperature}}{\underset{\text{diopside}}{2Ca(Mg, Fe)Si_2O_6} + \underset{\text{hypersthene}}{3(Mg, Fe)SiO_3} + \underset{\text{quartz}}{SiO_2}} . \tag{6}$$

The types of hornblende stable in the granulite facies are therefore rather rich in Al (the analyses available show from 8.65 per cent to 13.5 per cent Al_2O_3, according to Groves [1935] and my own unpublished data from Greenland).

We are dealing, therefore, with aluminous "common" hornblendes. The transition from the amphibolite facies to the granulite facies is, among other things, characterized by the following:

$$\overset{\text{low temperature}}{\underset{\text{hornblende}}{NaCa_2(Mg, Fe)_4Al_3Si_6O_{22}(OH)_2} + \underset{\text{garnet}}{(Mg, Fe)_3Al_2Si_3O_{12}} + \underset{\text{quartz}}{5SiO_2}} \rightleftharpoons \overset{\text{high temperature}}{\underset{\text{hypersthene}}{7(Mg, Fe)SiO_3} + \underset{\text{plagioclase}}{NaCa_2Al_5Si_7O_{24}}} . \tag{7}$$

The rhombic pyroxene-plagioclase side represents the granulite facies. Note that hornblende will be stable at higher *P-T* conditions in SiO_2-deficient rocks than in those which are quartz-bearing. In other words, the reaction

$$\overset{\text{low temperature}}{\underset{\text{hornblende}}{NaCa_2(Mg, Fe)_4Al_3Si_6O_{22}(OH)_2} + \underset{\text{garnet}}{(Mg, Fe)_3Al_2Si_3O_{12}} + \underset{\text{hypersthene}}{3(Mg, Fe)SiO_3}} \rightleftharpoons \overset{\text{high temperature}}{\underset{\text{olivine}}{5(Mg, Fe)_2SiO_4} + \underset{\text{plagioclase}}{NaCa_2Al_5Si_7O_{24}}} \tag{8}$$

takes place at higher *P-T* conditions than does reaction (7). Hornblende, rhombic pyroxene, anorthite or olivine, garnet, and/or spinel can therefore coexist in equilibrium in unsaturated rocks in the *P-T* interval between reactions (7) and (8). Therefore hornblende occurs more commonly in basic than in acid rocks in the granulite facies or in the charnockite series of rocks. Another cause of the abundance of hornblende in the basic members of the charnockite series is that an increase of anorthite in the plagioclase will displace equation (7) toward the hornblende side (see below).

However, it is evident that even in silica-saturated rocks equation (7) is not closely restricted to a given curve in the *P-T* field because the minerals taking part in the reactions are mixed crystals (Ramberg, 1944*c*).

The composition of the plagioclase is

important. Because hornblende is observed to have a greater Ca/Na ratio than the coexisting plagioclase, it is evident that an increase of the Ca/Na ratio in the system displaces reaction (7) toward hornblende, stabilizing this mineral at rising temperatures. In the Greenland rocks the plagioclase appears to be rather uniformly near An_{35} (tables 1, 2), so that, in the rocks concerned, reaction (7) would not be affected much by changing composition of plagioclase.

The Mn-concentration in the garnets in reaction (7) is also important. Principally, increasing Mn/(Fe, Mg) ratio, which stabilizes garnet, will displace the equilibrium toward the garnet side and, consequently, also toward the hornblende side. However, in most of the Greenland rocks biotite and potash feldspar occur with garnet so that under equilibrium conditions the spessartite content is kept constant at the same *P-T*. In harmony with this theoretical conclusion the garnet analyses (tables 5 and 6) show very little differences in the MnO content. For the Greenland rocks, therefore, reaction (7) will not be disturbed by changing the Mn/(Fe, Mg) ratio in the garnets.

The $Mg/(Fe^{\cdot\cdot\cdot}, Fe^{\cdot\cdot})$ ratio is more varied in the rocks in question, as is the content of TiO_2. It now appears that the Fe members in equation (7) must react at somewhat higher *P-T* conditions than the pure Mg members. This is supported by the distribution of Mg and Fe between coexisting hornblendes and pyroxenes (tables 1, 2, 3, and 4). Consequently, changing of the Fe/Mg ratio will influence the equilibrium represented by equation (7).

Analyses and optical investigations show that common hornblende of the type $(Na, Ca)_{2-3}$ $(Mg, Fe, Al, Ti)_5$ $(Si, Al)_8O_{22}$ $(OH, F)_2$ is considerably richer in TiO_2 than the coexisting rhombic pyroxene (Groves, 1935, p. 158, and unpublished data on the Greenland rocks). This means again that increasing Ti/(Mg, Fe) ratio in the rocks stabilizes hornblende at continuously rising temperature. However, the hornblendes from the granulite facies area in Greenland show rather constant content of TiO_2 ranging from about 2.4 to 4.2 per cent

In conclusion, we can construct a subsolidus equilibrium diagram (fig. 3, *A*) showing how a varying $Mg/(Fe^{\cdot\cdot}, Fe^{\cdot\cdot\cdot})$ ratio influences the stability of hornblende and rhombic pyroxene in the boundary field between granulite and amphibolite facies. In this diagram the hornblende curve and the hypersthene curve will be displaced upward if the plagioclase is more basic than An_{40}, or the percentage of TiO_2 in hornblende exceeds about 3-4 per cent. The equilibrium curves will be displaced downward if the plagioclase becomes more acid than An_{30}.

Furthermore, the rocks investigated have intermediate to large Fe/Mg ratios (tables 1-4), so that the correctness of the Mg side of the diagram is not verified. For example, one would expect cordierite to appear together with garnet or to replace garnet completely where the Mg/Fe ratio exceeds a certain limit (Eskola, 1939; Folinsbee, 1941, p. 501). It is also very likely that the reaction point on the Mg side of the diagram (point *B*, fig. 3, *A*) is situated so low that enstatite is replaced by anthophyllite or cummingtonite. Thus, the following reaction probably takes place where the pure Mg members are present:

$$\underset{\text{quartz}}{3SiO_2} + \underset{\text{cordierite}}{Mg_2Al_4Si_5O_{18}} + \underset{\text{hornblende}}{NaCa_2Mg_4Al_3Si_6O_{22}(OH)_2} \rightleftharpoons \underset{\text{anthophyllite}}{Mg_6Al_2Si_7O_{22}(OH)_2} + \underset{\text{plagioclase}}{NaCa_2Al_5Si_7O_{24}} \quad (9)$$

TABLE 1

QUARTZ-BEARING HORNBLENDE-ENDERBITIC GNEISSES FROM WESTERN GREENLAND*

		2,042	37,224	3,618	xHbg.	3,766	2,148	Hbg.4	Hbg.2	3,749	2,010	155	37,935	37,985	38,711
Plagioclase		An_{38}	An_{33}	An_{40}	An_{40}	An_{45}	An_{38}	An_{38}	An_{38}	An_{34}	An_{33}	An_{32}	An_{25}	An_{35}	An_{70}
Quartz		+	+	+	+	+	+	+	+	+	+	+	+	+	+
Potash feldspar		(+)	+	(+)	(+)	+	(+)	(+)	(+)	(+)	(+)	–	+	(+)	–
Rhombic pyroxene	γ	1.728	......	1.720	1.718	1.723	1.717	1.727	1.720	......	1.719	......	......	......	1.689
	α	......	–	......	......	......	......	......	......	+	......	–	–	+	......
	Mg/Fe	$\frac{52}{48}$	......	$\frac{58}{42}$	$\frac{59}{41}$	$\frac{56}{44}$	$\frac{61}{39}$	$\frac{53}{47}$	$\frac{58}{42}$	......	$\frac{59}{41}$	......	......	......	$\frac{82}{18}$
Monoclinic pyroxene	γ	1.723	......	......	1.716	......	......	1.723	......	......	......	......	......	......	......
	α	......	......	1.693	......	......	......	1.692	......	1.687	......	......	......	......	......
			–			+	–		–		–	+	–	–	–
Hornblende	γ	1.699	......	1.693	1.692	1.690	1.690	1.693	1.692	1.677	......	1.696	......	1.700	1.652
	α	1.677	1.692	1.673	β: 1.684	......	1.668	1.673	......	1.657	(+)	1.674	>1.698	1.678	1.632
	Mg/Fe	$\frac{40}{60}$	$\frac{20}{80}$	$\frac{47}{53}$	$\frac{48}{52}$	$\frac{50}{50}$	$\frac{50}{50}$	$\frac{47}{53}$	$\frac{48}{52}$	$\frac{65}{35}$	......	$\frac{43}{57}$	......	$\frac{40}{60}$	$\frac{77}{23}$
Biotite	β	......	......	......	......	......	1.643	......	......	......	1.643	......	......	......	......
	Mg/Fe	(+)	+	+	(+)	+	$\frac{45}{55}$	+	(+)	+	$\frac{45}{55}$	(+)	(+)	+	–
Garnet		–	+	+	–	–	–	–	–	–	–	–	+	–	–
Ilmenite and magnetite		+	+	+	+	+	+	+	+	+	+	+	+	(+)	(+)
Pyrite		–	–	–	(+)	–	–	–	+	+	+	+	+	+	(+)
Apatite		+	+	+	+	+	+	(+)	+	–	–	–	–	–	–
Sphene		–	–	–	–	–	–	–	–	–	–	–	+	–	–

---

* The relations between optical properties and chemical composition are taken from G. Kennedy, (1947); and S. Foslie, (1945). With reference to no. 38,711 it is doubtful that the rhombic pyroxene is stable in association with plagioclase. The Mg/Fe ratio in hornblende 38,711 is determined chemically.

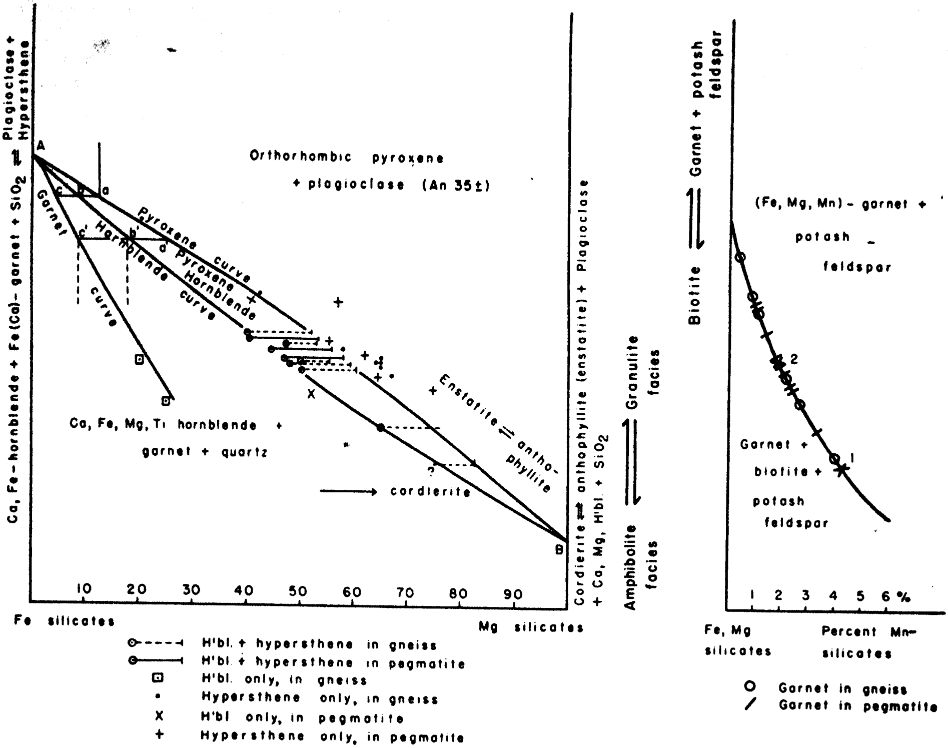

FIG. 3.—*A*, Subsolidus diagram of a saturated silicate system corresponding to composition given by equation (7), p. 36. The left-hand side of the diagram represents pure Fe components in this system, and the right-hand side represents pure Mg components in the same system. Point *A* corresponds to the transition point of the pure Fe members in equation (7), and point *B* to the transition point of equation (9).

Hornblende is unstable in the field: orthorhombic pyroxene –An_{35}. Rhombic pyroxene is unstable in the field: Ca, Fe, Mg, Ti-hornblende, garnet, quartz, cordierite. In the pyroxene-hornblende field are hornblende, rhombic pyroxene (and garnet plus plagioclase) in stable equilibrium.

Let us consider a paragenesis of An_{35} and hypersthene at decreasing temperature. At point *a*, for example, hypersthene begins to interact with plagioclase, forming hornblende of Mg/Fe ratio *b* and garnet of Mg/Fe ratio *c*. By decreasing temperature, the hypersthene follows the curve down to *a'*, the hornblende follows the curve down to *b'*, and garnet ends at *c'*. Now we suppose that at this point either hypersthene or plagioclase is completely used up in the reaction (eq. [7], p. 36) so that no more hornblende and garnet can form. Therefore, by further decrease in temperature, the composition of garnet and hornblende should be approximately constant following the stippled lines. (Because plagioclase is a mixed crystal, the last statement is not completely correct. However, a more thorough discussion is outside the aim of this paper.)

It should be noticed that the extremely Fe-rich and the most Mg-rich parts of the diagram are situated outside the actual composition of the investigated rocks. Nevertheless, the equilibrium curves are tentatively extended into these parts of the diagram. With reference to the Mg-rich part of the diagram we have some data which indicate that there is very little difference between the Mg/Fe ratios in hornblende and rhombic pyroxene.

B, This diagram represents the uppermost part of figure 2, *B*, thus being situated somewhat above the highest garnet marked in figure 2, *B*. The curve gives the spessartite content in garnets in equilibrium with biotite, ±sillimanite, and potash feldspar. All the garnets plotted are from the granulite facies area in western Greenland, tables 5 and 6. Point *1* represents garnet no. 37,817, table 6. Point *2* is the average per cent by weight of spessartite in the garnets from the granulite facies rocks.

Figures 3, *A*, and 3, *B*, are correlated to the transition interval between amphibolite facies and granulite facies and should therefore be easy to compare with figure 2, *B*.

Consequently, point *B* in figure 3, *A*, will be situated in the highest parts of amphibolite facies where enstatite is physically unstable. This would also harmonize best with the fact that cordierite very often is found in anthophyllite-bearing rocks but more seldom in regionally metamorphosed enstatite- or bronzite-bearing rocks.

Tables 1 and 2 give some mineralogical data from quartz-bearing hypersthene gneisses of the granulite facies area in western Greenland. According to the Mg/Fe ratio in the rhombic pyroxenes and the hornblendes in these rocks, they fall within a narrow *P-T* field in figure 3, *A*.

The MnO contents of garnets from some of the same gneisses and from khondalites, kinzigites, etc., in the same area are found in table 5 and plotted in the garnet-biotite equilibrium diagram, figure 3, *B*.

TABLE 2

QUARTZ-BEARING HORNBLENDE-FREE ENDERBITIC GENEISSES FROM WESTERN GREENLAND

	2,144	2,007	3,833	2,401	5,000	2,019	158	Hbg. 30
Plagioclase	An_{26}	An_{38}	An_{33}	An_{38}	An_{33}	An_{48}	An_{10}	An_{25}
Potash feldspar	(+)	(+)	(+)	(+)	(+)	(+)	(+)	++
Quartz	+	+	+	+	+	+	+	+
Rhombic pyroxene γ	1.720	1.712	....	1.712	1.712	1.717	....	....
α	....	....	1.700	....	1.700	....	1.696	1.728
Mg/Fe	$\frac{58}{42}$	$\frac{65}{35}$	$\frac{64}{36}$	$\frac{65}{35}$	$\frac{65}{35}$	$\frac{61}{39}$	$\frac{67}{33}$	$\frac{42}{58}$
Monoclinic pyroxene	+	–	–	–	–	+	–	+
Biotite β	....	....	....	1.637	1.638	....	....	....
Mg/Fe	+	–	–	$\frac{50}{50}$	$\frac{50}{50}$	+	+	(+)
Garnet	–	–	–	–	+	+	–	–
Ilmenite and magnetite	+	+	+	+	(+)	(+)	+	+
Pyrite	–	–	(+)	–	–	–	–	–
Apatite	+	+	+	+	+	(+)	+	+
Sphene	–	–	–	–	–	–	–	–

Table 3 shows some mineralogical data from the basic quartz-free rocks of the same district. It is seen that the Mg/Fe ratios in hornblende and rhombic pyroxene from these basic rocks are commonly greater than in the acid rocks. Thus, were we to place these minerals on the curves in figure 3, *A*, we would find that the basic rocks represent lower *P-T* conditions than the acid. However, in quartz-free rocks hornblende is stabilized owing to deficiency in SiO_2 (eqs. [7] and [8]), and the Mg/Fe ratio does not change with *P* and *T* in accordance with figure 3, *A*.

The hornblende-rhombic pyroxene assemblage in these basic rocks must be placed on the curves of figure 4. As is seen from the explanation of figure 4, the parageneses will in but few cases correspond to the olivine, hypersthene, enstatite, or hornblende curves where both olivine and plagioclase are present. Most of the mineral groups in table 3 are

TABLE 3

QUARTZ-FREE GNEISSES FROM WESTERN GREENLAND

		HYPERSTHENE-FREE ROCKS			HYPERSTHENE-BEARING ROCKS												
		3,739	2,038	2,075	2,188	2,074	2,016	2,129	3,310	Hbg.1	3,294	2,187	2,126	3,728	2,030	2,406	3,835
Plagioclase		An_{75}	An_{50}	An_{35}	An_{38}	An_{40}	An_{50}	An_{38}	An_{39}	An_{40}	An_{38}	An_{38}	An_{42}	An_{38}	—	An_{38}	An_{45}
Quartz		—	—	—	(+)	—	—	—	—	—	—	—	—	—	—	—	—
Potash feldspar		—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Rhombic pyroxene	γ				1.728	1.715	1.728		1.711		β: 1.700			1.711	β: 1.688	1.690	1.687
	α	—	—	—	1.708			1.703		1.700		1.692	1.688			1.677	1.675
	Mg/Fe				55/45	62/38	52/48	62/38	66/34	64/36	71/29	70/30	74/26	66/34	78/22	82/18	83/17
Monoclinic pyroxene	γ	1.727			1.715	1.715		1.720			1.716			1.711	1.713		
	α	1.703	1.703	+		1.688	—	1.695	+	1.690	1.688	1.690	1.683	1.687	1.687	—	1.677
Hornblende	γ	1.725	1.7105	1.692	β: 1.688	1.690	1.690	1.683	1.684	1.685	1.683	1.682	1.682	1.675	1.668	1.657	1.672
	α	1.703	1.6818	1.672		1.670	1.665	1.657	1.664	1.660		1.663	1.657	1.657	1.648	1.638	1.655
	Mg/Fe	?	46/54	47/53	40/60	50/50	53/47	62/38	58/42	59/41	60/40	60/40	62/38	67/33	74/26	?	69/31
Biotite	β				1.655					1.645	1.635	1.638		1.622			1.610
	Mg/Fe	—	—	+	30/70	—	—	+	+	45/55	50/50	48/52	+	60/40	—	+	70/30
Garnet		—	+	—	+	—	+	—	—	—	—	—	—	—	—	—	—
Olivine	γ														1.685		1.680
	Mg/Fe	—	—	—	—	—	—	—	—	—	—	—	—	—	92/8	—	95/5
Ore		+	+	+	+	+	+	+	+	+	+	+	+	+	Spinel	+	+

TABLE 4

MINERAL ASSOCIATIONS OF HYPERSTHENE-HORNBLENDE–BEARING PEGMATITES AND "APLITES" FROM WESTERN GREENLAND

		PEGMATITES									"APLITES"		
		3,302	2,001	2,409	3,276	160	112	2,039	37,218	360	3,307	Hbg.6	2,002
Plagioclase		An_{33}	An_{40}	+	An_{40}	+	An_{32}	An_{60}	An_{40}	An_{36}	An_{38}	An_{32}	An_{38}
Quartz		+	+	+	+	+	(+)	(+)	+	+	+	+	+
Potash feldspar		+	+	+	+	+	+	(+)	(+)	(+)	+	+	+
Rhombic pyroxene	γ	1.713	1.719	1.715	......	......	1.723	1.699	Anal.	1.720	......	......	......
	α	......	......	......	......	......	......	......	......	......	......	1.699	......
	Mg/Fe	$\frac{64}{36}$	$\frac{58}{42}$	$\frac{62}{38}$	(+)	−	$\frac{56}{44}$	$\frac{75}{25}$	+ $\frac{47.6}{52.4}$	$\frac{57}{43}$	+	$\frac{64}{36}$	+
Monoclinic pyroxene		−	−	−	−	(+)	−	−	+	−	+	α:1.687	+
Hornblende	γ	......	......	......	1.700	1.692	......	......	......	......	......	......	......
	α	......	......	......	1.673	......	......	......	......	......	......	......	......
	Mg/Fe	−	+	−	$\frac{40}{60}$	$\frac{48}{52}$	−	−	Anal. $\frac{44}{56}$	−	−	−	−
Biotite	β	......	......	......	......	......	......	1.625	......	1.650	......	......	......
	Mg/Fe	+	(+)	+	(+)	+	(+)	$\frac{60}{40}$	−	$\frac{40}{60}$	+	+	+
Ore		+	+	+	+	(−)	−	−	+	+	+	+	+
Apatite		+	(+)	(+)	(+)	+	−	+	(+)	(+)	+	+	+
Garnet		−	−	−	+	−	−	−	+	+	−	−	−

therefore situated between curve *Y* and the hornblende curve in figure 4. In this field hornblende and rhombic pyroxene exist in equilibrium, but only one or neither of the minerals olivine and plagioclase can be present.

We can conclude that the hornblende-hypersthene rocks of table 3 must lie within the same narrow *P-T* field as the assemblages of tables 1 and 2.

The harmony between field observation and mineralogical data from the granulite facies area in western Greenland is thus excellent; all data clearly show that the acid hypersthene gneisses and the garnet gneisses, as well as the basic noritic rocks, are formed or recrystallized at about the same pressures and temperatures.

Obviously, these different types of rocks did not crystallize out of a magma at the same *P-T* conditions. Such an as-

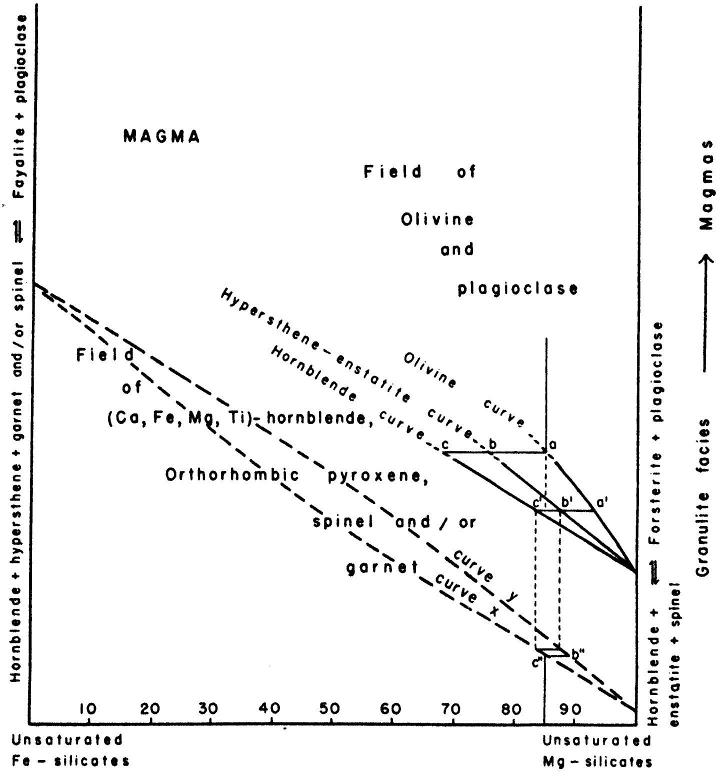
Hornblende + hypersthene + garnet and/or spinel ⇌ Fayalite + plagioclase
MAGMA
Field of Olivine and plagioclase
Hypersthene-enstatite curve
Hornblende curve
Olivine curve
Field of (Ca, Fe, Mg, Ti)-hornblende, Orthorhombic pyroxene, spinel and/or garnet
curve y
curve x
a
b
c
a'
b'
c'
b''
c''
10
20
30
40
50
60
70
80
90
Unsaturated Fe-silicates
Unsaturated Mg-silicates
Hornblende + enstatite + spinel ⇌ Forsterite + plagioclase
Granulite facies ——→ Magmas

FIG. 4.—Approximate equilibrium diagram for systems represented by equation (8), p. 36, in which Fe and Mg are mutually substitutive. The pyroxene curve and the hornblende curve in figure 3, *A*, are stippled in figure 4 to correlate the two diagrams.

At point *a*, plagioclase begins to interact with olivine, forming hypersthene of composition *b* and hornblende of composition *c* (plus spinel). By decreasing temperature, olivine follows the curve down to *a'*, and rhombic pyroxene and hornblende follow the curves to *b'* and *c'*, respectively. At that point, plagioclase and/or olivine might be completely used up in the reaction. By further decreasing temperature, therefore, the composition of the hornblende and rhombic pyroxene does not change until curve *y* is reached. At lower temperatures, rhombic pyroxene will disappear.

sumption is contrary to basic thermodynamics as well as to the empirically established reaction series of Bowen.

Only in a restricted and unlikely case can we assume that the lowest-melting silicate system among the gneisses happened to be in a liquid state during the metamorphism of the complex and that this magma then crystallized simultaneously as the rest of the complex underwent metamorphism and acquired stable mineral facies. In other words, the solidus temperature of the lowest-melting silicate systems would have had to coincide with the *P-T* conditions of the granulite facies. It follows, under this hypothesis, that the cross-cutting pegmatites (which in the magmatic theory are supposed to crystallize later and at lower temperatures than the massifs) should contain mineral suites indicative of a lower temperature than that of the regional metamorphic facies.

In the hypersthene gneisses there occurs a very characteristic type of pegmatite marked by the same brownish-gray antiperthite as the massive gneiss, the same fine perthitic potash feldspar as found in the charnockitic variety of the gneiss, and commonly the same bluish-milky or brownish quartz as that of the hypersthene gneisses. Moreover, hypersthene, a brownish-green hornblende, garnet, and biotite are common in these pegmatites.

The pegmatites are, as a rule, connected with small-grained cross-cutting dikes also having the same minerals as the enderbitic-charnockitic gneiss.

The mineralogical data from the farsunditic or charnockitic pegmatites and "aplites," encountered in the hypersthene gneiss, are shown in table 4 and plotted in figure 3, *A*. These data can be interpreted only by assuming that the hypersthene-bearing pegmatites and "aplites" are formed at the same *P-T* conditions as the enclosing gneisses.

In some places a somewhat different type of pegmatite occurs in the hypersthene gneisses. Pink potash feldspar and grayish-white oligoclase are characteristic of these pegmatites, which have only biotite and/or garnet as ferromagnesian minerals. In such mineral suites the MnO content in garnet is the best temperature index. Nos. 361, 388, 37,580, and 37,976 in table 6 are of this type, showing that these pegmatites likewise belong to the granulite facies.

No hypersthene occurs in the numerous cross-cutting and conformable pegmatites in the kinzigites, khondalites, and other hypersthene-free rocks intermingled with the hypersthene gneisses. Garnet, however, is abundant in the last-mentioned pegmatites. To compare the mineral facies of these pegmatites with that of the gneiss complex, a series of MnO determinations were made. Tables 5 and 6 give the results and show that the investigated pegmatites inside the granulite facies area all belong to the same high facies as the country gneisses and massifs.[5]

Thus, with the help, first, of the critical mineral rhombic pyroxene; second, of the subsolidus equilibrium between hornblende and pyroxene; and, finally, of the garnet, potash feldspar ⇄ biotite equilibrium, it is shown that the mineral as-

---

[5] It is noteworthy that the hypersthene richest in iron (no. *xyz*, table 5) is associated with the most Mn-poor garnet (fig. 3). Likewise, it is in complete harmony with the theoretical considerations that the garnet no. 37,817, table 6, with the highest content of spessartite found inside the granulite facies area, belongs to a pegmatite occurring only a few hundred feet from the rock no. 38,711, table 1. This rock, which is situated in the boundary zone between granulite-facies and amphibolite facies in Nagssugtoq, has a Mg-Fe distribution in its rhombic pyroxene and hornblende which places the paragenesis low in the hornblende—rhombic pyroxene diagram, figure 3, *A*.

TABLE 5

GARNET-BEARING GNIESSES FROM WESTERN GREENLAND*

	HOLSTEINSBORG			ISORTOQ		NAGSSUGTOQ				ARFERSIORFIK		NORTH OF AGTO
	3,286	2,188	2,024	37,547	xyz	37,612	37,887	37,787	37,847	37,390	37,380	37,935
Garnet % MnO % spessartite	1.16 2.69	0.85 1.97	0.4— 0.93	0.21 0.49	0.15 0.35	0.22 0.51	0.96 2.22	0.53 1.23	0.45 1.04	0.58 1.43	2.54 5.89	1.72 3.99 *n*: 1.799
Biotite β Mg/Fe	1.635 $\frac{5}{5}$	—	1.615 $\frac{7}{3}$	1.640 $\frac{5}{5}$	—	1.646 $\frac{4}{6}$	1.662 $\frac{2}{8}$	1.650 $\frac{4}{6}$	1.646 $\frac{4}{6}$	+	1.655 $\frac{3}{7}$	—
Plagioclase	+	+	+	+	+	+	+	+	+	+	+	+
Potash feldspar	+	—	+	(+)	(+)	(+)	+	—	+	—	—	+
Quartz	+	—	+	+	+	+	+	+	+	—	+	+
Others	Sillimanite	Hornblende Hypersthene Diopside	Sillimanite	Sillimanite	Hypersthene $\gamma \sim 1.77$ Fe/Mg = $\frac{80}{20}$	—	—	—	Cordierite	Cordierite	—	Hornblende Zircon Sphene

* All samples, except nos. 37,390 and 37,380, are from the granulite facies area (analysts: B. Bruun and H. Ramberg).

TABLE 6

GARNET-BEARING PREGMATITES FROM WESTERN GREENLAND*

		ISORTOQ						NAGSSUGTOQ					AGTO		ARFERSIORFIK			NORTH OF AGTO	
	361	342	343	388	37,512	37,580	360	37,738	37,718	37,817	37,848	37,857	37,976	37,218	37,364	37,381	37,423	37,878	37,449
Garnet																			
% MnO	0.50	0.75	0.61	0.44	0.95	1.07	0.39	0.79	0.43	1.83	0.73	0.51	0.99	0.82	1.43	7.38	20.0	4.46	3.66
% spessartite	1.16	1.74	1.41	1.02	2.20	2.48	0.91	1.83	1.00	4.25	1.69	1.13	2.29	1.92	3.32	17.12	46.0	10.35	8.49
n	......	......	1.788	......	1.788	1.788	......	1.788	1.783	......	1.785	......	1.783	......	1.798	......	......	1.80	......
Mg/Fe	......	......	$\frac{3}{7}$	......	$\frac{3}{7}$	$\frac{3}{7}$	......	$\frac{3}{7}$	$\frac{4}{6}$	......	$\frac{3}{7}$	......	$\frac{4}{6}$	......	$\frac{2\frac{1}{2}}{7\frac{1}{2}}$	......	......	$\frac{2\frac{1}{2}}{7\frac{1}{2}}$	......
Biotite β	......	1.650	1.650	1.652	1.653	1.648	1.650	1.654	1.635	1.640	1.648	1.650	1.646	......	1.646	......	......	1.670	......
Mg/Fe	+	$\frac{4}{6}$	$\frac{4}{6}$	$\frac{4}{6}$	$\frac{4}{6}$	$\frac{4}{6}$	$\frac{4}{6}$	$\frac{3}{7}$	$\frac{5}{5}$	$\frac{4\frac{1}{2}}{5\frac{1}{2}}$	$\frac{4}{6}$	$\frac{4}{6}$	$\frac{4\frac{1}{2}}{5\frac{1}{2}}$	—	$\frac{4}{6}$	+	—	$\frac{1\frac{1}{2}}{8\frac{1}{2}}$	—
Plagioclase	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+
Potash feldspar	+	+	+	+	+	+	(+)	+	+	+	+	+	+	(+)	+	+	+	+	+
Quartz	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+
Others	......	......	......	......	......	......	†	......	......	......	......	‡	......	§	‖	......	......	......	......

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* All samples are from the granulite facies area except nos. 37,381, 37,423, 37,878, and 37,449 (analyst: H. Ramberg).

† Rhombic pyroxene.

‡ Cordierite.

§ Hornblende, diopside, hypersthene.

‖ Sillimanite.

semblages of the pegmatites in the granulite facies in western Greenland are stable under granulite facies conditions. In addition, the comparatively low-temperature mineral muscovite, so commonly encountered in pegmatites in amphibolite facies areas, is found in only one of the hundreds of pegmatites examined. On the other hand, the high-temperature mineral sillimanite is encountered in some of the small conformable pegmatites penetrating the kinzigitic gneisses.

The chemical composition of these pegmatites in question exhibits considerable variation. No. 37,218, table 6, for example, is a plagioclase, hornblende, hypersthene, diopside, garnet, quartz, pegmatite with potash feldspar only in the form of antiperthite. No. 37,580, table 6, consists mostly of potash feldspar and quartz with garnet and biotite as the sole ferromagnesian minerals. No. 361 is a quartz vein with only small amounts of garnet, potash feldspar, and biotite. Therefore, the pegmatites in question cannot have the same melting interval; and, consequently, the only possible explanation is that they are formed during metamorphic-metasomatic processes below the solidus curve of silicate melts. Even without knowledge of the nature of these processes, it is evident that the granulite facies represents *P-T* conditions situated below the stability of "pegmatitic magma."

THE POSITION OF QUARTZO-FELDSPATHIC ROCKS IN THE MINERAL FACIES SCHEME

Unfortunately, no area has been studied from the viewpoint presented here, so it is difficult to compare the mineral facies of rocks and veins described in the literature. Therefore, let us first consider the facies classification of quartzo-feldspathic rocks, disregarding the degree of regional metamorphism in the area in which the rocks are situated.

For garnet-bearing quartzo-feldspathic rocks, the MnO content in the garnet is the most useful geothermometer. According to W. I. Wright (1938), the spessartite content in granite garnets ranges from almost 0 to 78.4 per cent and in pegmatite garnets from 1.4 to 90 per cent.

Placing these garnets in the equilibrium diagram, figure 2, *B*, we find that pegmatites and granites cover the same field of mineral facies ranging from low epidote-amphibolite facies up to granulite facies. (It has not been possible to determine whether the granite and pegmatite garnets of Wright's paper are in each instance in stable association with biotite. Some of the garnets may, therefore, have greater concentration of spessartite than that corresponding to the equilibrium curve in figure 2, *B*. This is true of the most spessartite-rich garnets recorded in Wright's paper.)

The epidote-plagioclase equilibrium in most quartzo-feldspathic rocks cannot register *P-T* conditions higher than those corresponding to An_{30-40} plus epidote, owing to the low Ca/Na ratio in these rocks. (Reference to Johannsen's descriptive petrography [1932, vol. 2] suffices to demonstrate that epidote or zoisite and An_{10-30} are commonly found in granodiorites, thus placing some of these rocks in epidote-amphibolite facies [fig. 2, *A*]).

In most pegmatites the Ca/Na ratio is still less than in massive rocks; therefore, the epidote-plagioclase equilibrium does not function, save below a rather low temperature limit. Examples of pegmatites with epidote and $An_{10\pm}$ are not rare (Andersen, 1931, p. 44; Erdmannsdörffer, 1931; and Wilkman, 1928).

Thus, the available data suggest that

most quartzo-feldspathic massifs and pegmatites exhibit mineral associations ranging from granulite facies down to epidote-amphibolite facies or actinolite greenschist facies. In other words, these rocks are formed and/or recrystallized below the field of magmatism.

Let us now briefly compare the mineral facies of various rocks and veins of areas outside the granulite facies.

AMPHIBOLITE FACIES AREAS

a) Some garnet-bearing pegmatites occur in the gneisses of the amphibolite facies north of the granulite facies complex in western Greenland. Nos. 37,449, 37,878, 37,423, and 37,381, table 6, are pegmatites from this area. No. 37,423 is a garnet replacing tourmaline in a biotite-free pegmatite. This garnet was formed, therefore, at higher temperature than that of the corresponding point on the equilibrium curve in figure 2, *B*.

Garnets from gneisses north of the granulite facies complex are shown in table 5, nos. 37,390 and 37,380. No. 37,935 is from the boundary zone between the granulite facies and the amphibolite facies. The rocks numbered 37,390 and 37,380 contain no potash feldspar, so the spessartite content in the garnets will be less than that corresponding to the equilibrium curve in figure 2, *B*.

However, the absence of hypersthene and sillimanite in this area shows that the gneisses around the pegmatites correspond to the amphibolite facies.

TABLE 7

ROCK TYPES	INDEX MINERALS FROM ROCKS OF KANTORP AREA, SWEDEN*			INDEX MINERALS FROM ROCKS OF NAMAQUALAND, AFRICA†		
	Old Gneiss, Leptite	Younger Granites	Pegmatites	Old Gneiss	Younger Granites	Pegmatites
Plagioclase	An_{8-70}	$An_{25}\pm$	Oligoclase	Andesine	An_{10-43}	+
Potash feldspar	+	+	+	+	+	+
Biotite	+	+	+	+	+	+
Sillimanite	+	+	+	−	−	−
Andalusite	+	+	+	−	−	−
Cordierite	+	+	+	−	−	−
Garnet	+	+	+	−	−	Spessartite
Epidote	−	−	−	+	+	+
Chlorite	−	−	−	+	+	+
Muscovite	?	?	?	+	+	+
Hypersthene	−	−	−	−	−	−
Anthophyllite	(+)	−	−	−	−	−
Sphene	±	±	?	+	+	+

* After Magnusson. † After Coetzee; Gevers, Patridge, and Joubert.

b) The Kantorp ore area in Sweden is worthy of special mention. According to Magnusson's (1936) description, the rocks have been formed and recrystallized under rather constant *P-T* conditions corresponding to the amphibolite facies. The index minerals are cordierite, sillimanite, andalusite, and anthophyllite. The highest *P-T* limit is indicated by the absence of hypersthene and the lower by the absence of epidote in spite of very basic plagioclase (up to An_{70}).

Table 7 shows the pertinent assemblages in the old gneisses, the younger granites, and the pegmatites in the Kantorp area. The whole complex, including both the conformable and the cross-cutting pegmatites, must have been formed at nearly the same *P-T* conditions, distinctly below the granulite facies, but well above the epidote-amphibolite facies.

c) Around Söndeled in southern Norway occur rocks belonging to the amphibolite facies and containing cordierite and anthophyllite but neither hypersthene nor epidote (Andersen, 1931, p. 43). Cross-cutting and conformable pegmatites carry the same index minerals (anthophyllite and cordierite) as the host gneisses.

d) A. Brammall and S. Bracewell (1936) have studied the garnets in the well-known Dartmoor granite. Muscovite, andalusite, and cordierite indicate approximately the facies. These minerals plus the absence of epidote place the granite in the amphibolite facies and distinctly below the granulite and pyroxene-hornfels facies.

Analyses of fourteen garnets from localities in the granite massif reveal great variation in manganese content, ranging from 3.3 per cent MnO (7.82 per cent by weight of spessartite) to 22 per cent MnO corresponding to 51 per cent spessartite. Thus, none of these garnets is so poor in spessartite as those from the rocks of the granulite facies in Greenland.

Brammall and Bracewell (1936) did not list the mineral associations of garnets with more than 7.3 per cent MnO, so the presence of biotite is in doubt. Therefore, it is not certain that the assemblages with the most spessartite-rich garnets actually represent as low temperatures as indicated by the Mn/(Fe, Mg) ratio. The garnets with 7.82–17 per cent by weight of spessartite are associated with both biotite and potash feldspar, making these garnets usable facies indicators for the Dartmoor granite. Probably the Dartmoor granite was formed, or recrystallized, at lower temperatures than the pegmatites from the granulite-facies area in Greenland.

Cordierite occurs with garnets of 7.30 per cent MnO (or 17 per cent spessartite), showing that cordierite is stable down to comparatively low temperatures.

EPIDOTE-AMPHIBOLITE FACIES AREAS

The literature contains insufficient data regarding this facies, so only a few examples can be considered.

a) South of the Orange River in Namaqualand in South Africa is an area of old gneisses and schists passing into younger granites to the south and east. The boundary between the old and young complexes is not sharp (Gevers, Patridge, and Joubert, 1937; Coetzee, 1941).

Coetzee's description shows that the old gneisses correspond partly to the highest epidote-amphibolite facies and partly to the lowest amphibolite facies. The younger granites to the south and east belong to a somewhat lower facies apparently not exceeding the highest boundary of the epidote-amphibolite facies. Table 7 shows the plagioclase-epidote associations in rocks described by Coetzee.

Whereas in western Greenland the gneissification and pegmatitization have taken place in the granulite facies, the activation in Namaqualand has taken place in the epidote-amphibolite facies and contains minerals consistent with this low-grade metamorphism. The Namaqualand pegmatites are not recorded as containing sillimanite, cordierite, or hypersthene but do carry muscovite and epidote (Gevers, Patridge, and Joubert, 1937).

The common garnet in the pegmatites is of rather constant composition, with the index of refraction ranging only from

1.815 to 1.820 and the density from 4.12 to 4.213 (Gevers, Patridge, and Joubert, 1937, p. 101). Analysis of one garnet shows 26.96 per cent MnO, corresponding to 61.55 per cent spessartite.

Placing this garnet in the equilibrium diagram, figure 2, *B*, reveals that it agrees closely with the epidote-plagioclase equilibrium of the host gneisses in the area.

b) In central Finland, east of Oulainen, many occurrences of unakite are reported (Wilkman, 1928). That the mineral assemblage of these masses belongs to the lowest epidote-amphibolite facies is indicated by the association of epidote and albite. In these rocks "primary" pegmatites are encountered with microcline, acid albite (An_{3-5}), epidote, quartz, chlorite, hematite, and apatite.

c) In the neighborhood of Notodden, in southern Norway, lies a body of epidote-oligoclase granite (Andersen, 1931, p. 44). Pegmatites in the granite contain epidote and acid plagioclase of nearly the same composition as the granite.

Andersen states that epidote was the first mineral to crystallize in the pegmatites, thus proving that epidote in this case cannot be interpreted as a "secondary" product.

d) At Weissenstein near Stammbach, Germany, occurs a zoisite-bearing pegmatite with plagioclase (An_{13-16}), muscovite, quartz, and some small garnets. According to O. H. Erdmannsdörffer (1931), the pegmatite is enclosed in a garnet-zoisite amphibolite with the same plagioclase as is found in the pegmatite.

e) Garnets from the pegmatites of southern Norway were investigated by the writer some years ago. The pegmatites are found in various rocks belonging to the epidote-amphibolite facies and to the amphibolite facies. The pegmatite garnets contain about 20–82 per cent by weight of spessartite. However, lack of samples of country rock associated with the pegmatites prevented direct comparison of the mineral facies.

CONCLUSIONS REGARDING THE FORMATION OF PEGMATITES

It is concluded that most pegmatites and quartzo-feldspathic masses in folded mountain zones were formed below the solidus curve of any magma. Furthermore, the pegmatites and most masses of quartzo-feldspathic composition have mineral assemblages of substantially the same facies as the country rocks. Therefore, it is concluded that most pegmatites and acid massifs in regional metamorphic areas were formed with the help of some submagmatic metamorphic-metasomatic processes.

Subsolidus formation of pegmatites and massifs takes place over a very wide *P-T* field: the epidote-amphibolite facies, the amphibolite facies, and the granulite facies. Only beyond the granulite facies at high pressure and in, or above, the pyroxene hornfels facies at low pressure can rock-making silicate melts exist. A discussion of the forces and kinetics underlying the metamorphic and metasomatic processes is not the aim of this paper.

Metasomatic-metamorphic pegmatites can be most conveniently classified into four groups:

1. *Secretion pegmatites* are formed by diffusion of atoms, molecules, and ions through the rocks toward open fissures and other places where mechanical pressure is low. Owing to the low mechanical pressure in such openings, the vapor tension of the minerals which grow here will also be low. Consequently, there exist forces which tend to make the activated movable atoms, ions, and molecules diffuse toward the openings where the ac-

tivated particles must consolidate into minerals.

2. *Concretion pegmatites* grow in solid rocks during regional metamorphism and push the country rocks in front of their expanding boundaries, thus developing a conformably stressed external contact zone. Because the mechanical pressure is high in the growing pegmatite and low in the surroundings, this type of petroblastic rock cannot be due to chemical potential differences induced by pressure differences. The driving agency here is the influence of the environment on the chemical potential or the vapor tension of the minerals. That is, feldspar and quartz intermingled with basic minerals should have a higher vapor tension than feldspar and quartz alone. Thus the diffusion processes tend to form concretions of more or less pure quartzo-feldspathic bodies.

3. Whereas in 1 and 2 the substances supplied might consolidate without reaction with old minerals in the space where the pegmatites form, *replacement pegmatites* are formed by intense reaction between introduced and pre-existing substances. The chemical potential differences necessary to form these pegmatites—to make the substances migrate toward the growing pegmatite and to make the activated particles consolidate—must be due to chemical incompatibility between the pre-existing minerals and the surrounding rocks.

4. *Recrystallization pegmatites* are formed by coarse-grained metamorphic recrystallization within limited spaces in the rocks. Pegmatites formed in this way will have almost the same chemical and mineralogical composition as the enclosing gneisses. The conformable cordierite-garnet pegmatites in kinzigitic gneisses are supposed to belong to this type.

Pegmatites are commonly of quartzo-feldspathic composition regardless of the composition of the country rocks. This is not because a quartz-feldspar system has the lowest melting temperature but because the elementary constituents of quartz, potash feldspar, and acid plagioclase are able to diffuse more readily than the constituents of most basic minerals. However, basic constituents can diffuse to some degree, and, therefore, the minor minerals in pegmatites (biotite, garnet, diopside, hornblende, cordierite, and several of the rare minerals) actually depend on the type and composition of the country rocks.

Only where the country rock is very poor in potash, soda, silica, etc., and the sources of these substances are distant will the pegmatites have a nonquartzo-feldspathic composition. Good examples are the corundum-bearing pegmatites found in ultrabasic rocks in the northern Transvaal (Brandt, 1946).

The chemical interrelation between pegmatites and the country rocks requires much more investigation. One interesting point is that many pegmatites can be classified according to content of *groups* of elements. Thus the Li pegmatites are rich not only in Li but also in other typical elements accompanying Li, such as Sn and F. This grouping of elements seems to be easily explained if it is assumed that the substances are introduced as liquid solutions or silicate melts charged with the typical group of elements required.

However, if the elements migrate in the form of individual atoms, molecules, or ions, a certain interdependence and grouping of the elements must also be expected. The principal reason is that migrating atoms, ions, or simple molecules cannot usually consolidate into crystalline elements or simple molecules in the pegmatites. The vapor pressure of

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such crystalline elements or simple molecular compounds is so high that the elements will tend strongly to combine into more complicated compounds having lower partial vapor pressure of their several constituents. Thus, owing to chemical forces, certain groups of elements will tend to consolidate in pegmatites despite the more or less independent migration of these elements toward the growing pegmatite. In micas, for example, there is a relationship between the content of Li and the F/OH ratio so that Li-rich micas tend to be rich in F^- and poor in OH^- because of interionic polarization forces.[6]

In two hypothetical rock complexes, assume the same small concentration of Li_2O but different concentrations of F. In both complexes pegmatites form during metamorphism. It is likely that the pegmatites growing in the F-rich rocks will favor the formation of Li micas, whereas Li does not find conditions favorable for consolidation into Li minerals in the pegmatites forming in the F-poor complex.

Similarly, if both rock complexes contain traces of Sn, cassiterite will most likely form in the Li pegmatites in the F-rich complex but not in the pegmatites in the F-free complex. This is because the volatile SnF_4 molecules will diffuse more easily intergranularly than individual Sn atoms, ions, or less volatile Sn compounds.[7]

Because we must doubt the magmatic origin of most pegmatites and even the importance of circulating pore solutions in metamorphic-metasomatic rocks, we have to look for forces and processes of the kinds indicated above in our speculations about the formation of pegmatites and other bodies of quartzo-feldspathic composition.[8]

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[6] According to oral communication from Dr. F. Wickman.

[7] On field and experimental evidences as well as on theoretical grounds the writer considers it unlikely that large-scale diffusion through solid rocks preferably takes place through the silicate lattices. Many large atoms or ions (K, F, etc.), and also molecules such as H_2O, CO_2, etc., are much more mobile than are small ions or atoms. Furthermore, it can be shown that the mobility of the several particles in rocks increases with rising partial vapor pressure of the particles. It seems very probable that large-scale migration through solid rocks takes place with the help of "gaseous" atoms and molecules adhered to the mineral surfaces or the mosaic fissures. Only to a considerably less extent will lattice diffusion extend over large distances.

[8] After this paper was sent to press, garnets from two new pegmatites from the amphibolite and epidote-amphibolite facies areas in western Greenland were analyzed. No. 36647 contains 12.45 per cent MnO; No. 12020 contains 19.22 per cent MnO.

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FREQUENCY DISTRIBUTION OF THE MINERALS IN TWO PETROGRAPHIC PROVINCES[1]

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ABSTRACT

The frequency distribution of the major rock-forming minerals in the Oslo petrographic province and in the intra-Pacific province has been determined. In a concluding section several ways to arrive at the average mineral composition of igneous rocks are presented for consideration.

INTRODUCTION

Petrographers have noticed as a significant and striking fact that, although more than a thousand minerals are known, the number that composes over 99 per cent of igneous and metamorphic rocks is very small. Indeed, minerals that are really essential in rock-forming processes number somewhat less than a score, depending upon how strictly each mineral species is defined.

Recent trends in petrography have been toward quantitative calculations; although numerous detailed rock descriptions, including modes, have been published, no quantitative information on the regional frequency distribution of the several minerals is available.

In this paper mineral data on the igneous rocks of the Oslo province and of the intra-Pacific province are presented and statistically treated. Petrochemical studies of these provinces, published long ago, should be consulted for detailed descriptions (Clarke and Washington, 1924; Daly, 1944; Barth, 1931 and 1945; Macdonald, 1946 and 1947).

METHODS OF CALCULATION

A quantitative mineral determination in a rock can be made with a microscope and an integrating stage. This is known as the "Rosiwal analysis." The process is often tiresome and may require specially prepared thin-sections. The recalculation of chemical data is more convenient if large numbers of chemical analyses are available from a rock province.

In such calculations molecular (or atomic) proportions must be used. Here I want to repeat my old plea to all petrographers that the results also be stated in terms of molecules. There is no special significance attached to the weight of a mineral, and I see no reason whatsoever for converting the molecular proportions into weight percentages. The chemical relations of a rock are obscured by the use of weight percentages; moreover, the computations become unnecessarily cumbersome.

It seems nonsensical to me that Billings and Rabbitt (1947) should write:

> Although the use of molecular proportions greatly facilitates the calculations, it is less precise than it would be to assign the oxides by weights per cent. . . . Niggli (1936) and Barth (1945) have advocated using atomic percentages in all petrographic calculations. Whatever the advantages of Niggli's system, the "slack" is even greater than in the method employing molecular proportions.

This statement is equivalent to saying that, although the use of the metric system facilitates most calculations,

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[1] Manuscript received June 7, 1948.

measurements taken with a meter stick are less precise than those taken with a yardstick. Furthermore, Billings and Rabbitt, as well as everybody else who wants to present a rock analysis in terms of mineral molecules, use molecular (or atomic) proportions as one step in the procedure. The results thus obtained cannot possibly become more accurate by being restated in terms of weight percentages.

The fundamental question, stripped of all formalities, is whether or not it is possible to calculate the mineral composition with no other knowledge than the chemical analysis of an igneous rock. The answer, in the general case, is that it is not possible. There are many reasons for this, the most important probably being that the crystallization (or recrystallization) of most rocks took place under changing P-T conditions which prevented establishment of equilibrium among the minerals; zoned feldspars, armored relics, and glass attest to disequilibria. Obviously, therefore, the mineral composition cannot be regarded as a one-valued function of the chemical composition, even within a defined mineral facies. Nevertheless, in most rocks the approach to equilibrium has limited the possibilities of variation in the mineral assemblages to such an extent that, with some additional information, accurate results can be obtained.

In most cases it will be necessary to know the chemical composition or the amount of one or more of the minerals present. Still, there is no automatic approach, as in the case of the norm calculations; but trial-and-error methods must be employed. It becomes of great importance, therefore, to use a quick and simple method of computation; otherwise the expenditure of time becomes prohibitive. The best method for this purpose is that proposed by Niggli in 1936.

In petrographic calculations it has been customary to use the proportions of the molecules, such as SiO_2, Al_2O_3, etc. Niggli recommended the use of the proportions of the several cations (silicon and metals); recalculated to 100, they represented what Niggli called the "equivalent molecular percentages" which form the basis for the calculation of the mineral molecules (table 1). It should be noticed that in this scheme the molecular amount of a mineral becomes propor-

TABLE 1

CALCULATION OF THE MINERAL CONTENTS OF AN OSLO-ESSEXITE FROM RANDVIKSHOLMEN

	Si	Ti	Al	$Fe^{\cdot\cdot\cdot}$	$Fe^{\cdot\cdot}$	Mg	Ca	Na	K	P	Sum
ATOM PER CENT OF ROCK	45.7	1.4	14.4	5.2	6.6	8.1	9.4	6.7	2.1	0.4	100.0
Alkalifeldspar ($Or_{60}Ab_{40}$)	7.5	......	2.5	......	......	......	......	1.0	1.5	......	12.5
Plagioclase (An_{34})......	20.5	......	10.3	......	......	......	2.6	5.1	......	......	38.5
Pyroxene............	15.5	0.9	1.4	1.2	2.9	6.4	6.0	0.6	0.2	......	35.1
Olivine (Fa_{33}).........	1.1	......	......	......	0.7	1.5	......	......	......	......	3.3
Biotite..............	1.1	0.1	0.2	0.4	0.7	0.2	......	......	0.4	......	3.1
Ore...................	......	0.4	......	3.6	2.2	......	......	......	......	......	6.2
Apatite..............	......	......	......	......	......	......	0.6	......	......	0.4	1.0
Sum............	45.7	1.4	14.4	5.2	6.5	8.1	9.2	6.7	2.1	0.4	99.7

tional to the sum of the cations in the mineral. In this way the mineral equations become very simple, for instance:

$$\begin{array}{ccccc} NaAlSiO_4 & + & 2SiO_2 & = & NaAlSi_3O_8 \\ 3Ne & + & 2Q & = & 5Ab \end{array}$$

(the nepheline formula contains three cations, the quartz one cation, and the albite five cations).

An example of this calculation of the mineral composition of an alkali gabbro is given in table 1. In this case the chemical composition of the constituent pyroxene was known (and the nature of the feldspars was determined optically). In other cases, if no mineral analysis was available, the amounts of one or several of the minor minerals were determined by Rosiwal analyses. In no cases do the calculations become wholly automatic, however; recourse must be had to trial-and-error methods. Nevertheless, the results are very reliable. One cannot fail to become impressed by the fact that a few simple and routine observations with the microscope usually limit the possibilities so much that the results are very nearly correct. For further details, the reader is referred to the previous papers by Niggli and Barth, cited above.

THE OSLO PETROGRAPHIC PROVINCE

Through the classical works of the last century by Kjerulf and Brögger, the Oslo area became one of the best-known rock provinces. The Oslo plutonics form a mildly alkaline series, ranging from alkali gabbro through syenite (and nepheline syenite) into ægerite granite (and biotite granite). Brögger regarded these rocks as comagmatic and genetically related to the corresponding lavas. The areal distribution of the principal rock types in the Oslo province is listed in table 2, which is quoted from an earlier publication (Barth, 1945).

The same paper contains data for the quantitative mineral composition of sixty-four plutonic igneous rocks. Additional data are found in papers by Sæther (1945) and by Oftedahl (1946). These data are treated here statistically, and the results are exhibited in tables 3–9.

TABLE 2

AREAL DISTRIBUTION OF THE PRINCIPAL ROCK TYPES IN THE OSLO PROVINCE

Rock Type	Km.²	Per Cent
Oslo-essexite	15.3	0.3
Akerite	52.2	1.0
Kjelsasite (= mica syenite, in part)	201.0	4.0
Lardalite (= nepheline syenite)	65.0	1.3
Larvikite (= augite syenite)	1670.0	32.8
Nordmarkite-pulaskite	1425.0	28.0
Ekerite (= soda granite)	821.0	16.1
Biotite granite	840.0	16.5
Plutonic rocks	5089.5	100.0
Basaltic lava	220.0	15.6
Rhomb porphyry	1160.0	82.2
Quartz porphyry	30.6	2.2
Extrusive rocks	1410.6	100.0
	6500.1	....

THE INTRA-PACIFIC ROCK PROVINCE

The rocks of the intra-Pacific islands are preponderantly of basaltic habit, with small amounts (*ca.* 3–5 per cent) of alkaline lavas, mostly trachytes and phonolites. The mean mineral composition of these two rock groups is shown in table 10. The rocks are well known both chemically and mineralogically (Macdonald, 1947); detailed petrographic descriptions, including modes calculated in weight per cent, have been given for sixty rock specimens (Barth, 1931). These investigations form the basis of the present statistical studies.

If we assume the relative areal distribution of basaltic rocks, on the one side, and phonolite-trachytic rocks, on the other, to be 95 per cent and 5 per cent, respectively,[a] the average mineral composition of the whole petrographic province comes out as shown in column III, table 10.

TABLE 3

AVERAGE MINERAL COMPOSITION OF ESSEXITES AND BASALTS

	I*	II†
Plagioclase	40	47
Augite	23	20
Hornblende	8.5	10
Alkali feldspar	7.5	8
Ores	7.5	8
Biotite	5.5	...
Olivine	3.0	...
Serpentine	2.0	2
Apatite	1.0	...
Sphene	0.5	0.5
Calcite	0.5	1
Quartz	0.4	1
Chlorite	0.3	2
Hypersthene	0.2	...
Muscovite	0.1	...
Epidote	...	0.2
Glass	...	0.2
Zeolite	...	0.1
Total	100.0	100.0

* Twenty Oslo-essexites. † Four Oslo-basalts.

TABLE 4

AVERAGE MINERAL COMPOSITION OF KJELSASITES AND AKERITES

	I*	II†
Plagioclase	42.0	39
Alkali feldspar	28.5	35
Quartz	6.0	6
Augite	6.0	2
Hornblende	6.0	10
Ores	5.0	3
Biotite	3.0	3
Apatite	2.0	0.7
Serpentine	0.8	1
Olivine	0.4	...
Hypersthene	0.3	...
Sphene	0.03	0.3
Calcite	0.03	...
Total	100.06	100.0

* Seven kjelasites. † Ten akerites.

COMPARISONS AND CONCLUSIONS

It stands to reason that the average mineral composition of the whole intra-Pacific province cannot be much different from that of the Hawaiian rocks.

Daly (1944) has computed the average chemical composition of a "primary oli-

TABLE 5

AVERAGE MINERAL COMPOSITION OF TWO LARDALITES

Aklaki feldspar	54
Nepheline	16
Ores	10
Pyroxene	7
Biotite	5
Olivine	4
Apatite	2.5
Sodalite	1.0
Hornblende	0.5
Total	100.0

TABLE 6

AVERAGE MINERAL COMPOSITION OF LARVIKITES, RHOMB PORPHYRIES, AND NORDMARKITES

	I*	II†	III‡
Alkali feldspar	45.5	60	87
Plagioclase	34.0	25	...
Augite	7.5	2	1
Hornblende	2.5	...	3
Ores	2.5	1	2
Quartz	2.0	3	2
Biotite	2.0	...	2
Pyroxene	1.0§	...	1.5‖
Olivine	0.9	...	...
Apatite	0.9	0.5	0.4
Nepheline	0.7	...	...
Sphene	0.4	...	1
Calcite	0.1	2	0.1
Chlorite	...	5	...
Muscovite	...	1	...
Epidote	...	0.5	...
Total	100.0	100.0	100.0

* Seven larvikites

† Sixteen rhomb porphyries.

‡ Eight nordmarkites.

§ Hypersthene.

‖ Ægerite.

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[a] This may be an overestimation of the amounts of alkaline lavas; but, if one takes only half this amount, the results will not be notably changed.

TABLE 7

AVERAGE MINERAL COMPOSITION OF EKERITES, GRANITES, AND QUARTZ PORPHYRIES

	I*	II†	III‡
Alkali feldspar	73.5	70.5	60.0
Quartz	18.5	26.0	36.0
Ægerite	4.5	....	....
Ores	1.2	1.5	1.5
Biotite	....	0.7	1.0
Plagioclase	....	....	0.5
Sphene	1.0	0.4	0.4
Hornblende	1.0	0.3	0.2
Chlorite	....	0.3	0.3
Apatite	0.1	0.2	0.1
Zircon	0.1	....	....
Fluorite	0.1	0.1	....
Total	100.0	100.0	100.0

* Six ekerites. In the ekerites, elpidite and astrophyllite are present in amounts less than 0.1 per cent.

† Four granites.

‡ Two quartz porphyries.

TABLE 8

WEIGHTED AVERAGE MINERAL COMPOSITION OF OSLO ROCKS

	I*	II†
Alkali feldspar	52	65
Plagioclase	28	13
Quartz	3	9
Pyroxene	5	5‡
Chlorite	4	....
Ores	2	2
Hornblende	2	2
Calcite	2	....
Biotite	....	2
Muscovite	0.8	....
Apatite	0.4	0.6
Epidote	0.3	....
Olivine	0.3§	0.4
Sphene	0.1	0.6
Nepheline	....	0.4
Inclusive	0.1‖	¶
Total	100.0	100.0

* All extrusive rocks.

† All plutonic rocks.

‡ Including 1.5 per cent ægerite, 0.3 per cent hypersthene, 3.2 per cent augite.

§ Serpentine.

‖ Present in amounts <0.1 per cent: glass, biotite, zeolite.

¶ Present in amounts <0.1 per cent: chlorite, serpentine, calcite; <0.01 per cent: fluorite, sodalite, zircon, astrophyllite, elpidite.

TABLE 9*

WEIGHTED AVERAGE MINERAL COMPOSITION OF ALL OSLO ERUPTIVE ROCKS (86 ANALYSES)

(Surface Area = 6500 Km.²)

Aklaki feldspar	62
Plagioclase	16
Quartz	8
Pyroxene	5
Hornblende	2
Ores	2
Biotite	2
Chlorite	0.8
Sphene	0.5
Apatite	0.5
Calcite	0.4
Nepheline	0.3
Olivine	0.3
Muscovite	0.2
Total	100.0

* Present in amounts <0.1 per cent: serpentine and epidote; <0.01 per cent: fluorite, sodalite, zircon, glass, zeolite, astrophyllite, and elpidite.

TABLE 10

AVERAGE MINERAL COMPOSITION OF PACIFIC ROCKS

	I*	II†	III‡
Alkali feldspar	10	61	13
Plagioclase	31	18	30
Pyroxene	28	6§	27
Olivine	10	0.2	9.5
Ores	7.5	5	7.4
Glass	7.7	2	7.4
Hornblende	....	2	0.1
Quartz	1.0	2.5‖	1.1
Nepheline	2.3	1.6¶	2.2
Apatite	1.2	0.6	1.2
Serpentine, etc.	0.5	0.5	0.4
Biotite	....	0.5	....
Inclusive	0.8	0.1	0.7**
Total	100.0	100.0	100.0

* Fifty rocks of basaltic habit; includes twenty-seven olivine basalts, twelve basalts, five pacificites, five tephrites and basanites, and one trachybasalt.

† Two phonolites + two trachytes + one rhyolite.

‡ All Pacific lavas.

§ Including 1.4 per cent acmite.

‖ Including 0.6 per cent tridymite.

¶ Including 0.5 per cent analcite.

** Present in amounts less than 0.1 per cent are biotite, tridymite, cristobalite, epidote, prehnite, calcite, melilite, sodalite, noselite, hauyne, sphene, zircon, kaolinite, sulfur, and sulfates.

vine basalt," which he regards as the material out of which all the lavas of Hawaii have been made, and for the most part without contamination. However, some of the alkaline lavas (nepheline basalt, etc.), he believes, were formed through reaction with carbonate rocks.

If the primary olivine basalt represents the mother magma, then it should be capable of producing a mineral assemblage similar to that listed in column III, table 10, for the Pacific province.

TABLE 11

MINERAL COMPOSITION OF PACIFIC LAVAS

	I*	II†	III‡
Alkali feldspar	18	9	15
Plagioclase	30	40	34
Pyroxene	27	34	26.5
Olivine	9.5	5.7	10.5
Ores	7.4	6.6	6.1
Quartz	3.5	3.5	3.5
Nepheline	2.2	0.0	2.7
All others	2.4	1.2	1.7
Total	100.0	100.0	100.0

* Average mineral composition of all Pacific lavas, quoted from column I, table 10. The glass which is present to the amount of 7.4 per cent in the Pacific lavas has, for comparison purposes, been recalculated into 2.4 per cent quartz and 5 per cent alkali feldspar.

† Computed mineral composition of Daly's "primary olivine basalt."

‡ Computed mineral composition of all Hawaiian lavas from Washington's average chemical analysis.

Table 11 shows that Daly's primary olivine basalt magma cannot produce such a mineral assemblage: alkali feldspars are too low, plagioclases are too high, and, if quartz is to correspond to the amounts in which it is actually observed, then the computed amounts of the undersaturated minerals (olivine and nepheline) become too low.

This is what one would expect; for Daly believes that some of the "alkaline" rocks of Hawaii were formed through assimilation of limestone, the obvious result of which will be to increase the amounts of olivine and nepheline.[3]

Instead of Daly's primary magma, we will, therefore, consider the average chemical composition of all Hawaiian lavas, as determined by Washington, in which he took the average of all good analyses of such lavas. This average analysis can be recalculated into a mineral assemblage which comes very close to the one observed for the Pacific province (table 11).

The mineral calculations referred to above can be checked easily by the data displayed in table 12. The assumptions involved are that the pyroxene phases contain small amounts of Ti, Al, and ferric iron (from 0.5 to 2 per cent), and the composition of the alkali feldspar is 40 Or, 60 Ab.

There are no accurate data on record of the average mineral composition of igneous rocks; but much thought has been given to the chemical composition. By taking the average of all good analyses, Clarke and Washington (1924)

TABLE 12

CHEMICAL ANALYSES STATED IN TERMS OF PERCENTAGES OF THE CONSTITUENT CATIONS

	I*	II†	III‡
Si	47.0	46.4	55.8
Ti	2.2	1.9	0.8
Al	14.9	15.0	17.1
Fe^{+3}	1.9	2.0	2.2
$Mn^{+2}+Fe^{+2}$	7.2	6.8	3.0
Mg	10.9	11.5	4.9
Ca	10.6	9.1	5.1
Na	4.4	5.7	7.1
K	0.7	1.2	3.8
P	0.2	0.4	0.2
Total	100.0	100.0	100.0

* Primary olivine basalt of Hawaii (Daly's average).

† All lavas of Hawaii (Washington's average).

‡ All igneous rocks (Clarke and Washington's average).

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[3] But not of alkali feldspar. Therefore, the writer believes in a more alkaline primary magma rather than in limestone syntexis.

have arrived at figures which are generally believed to represent the average chemical composition of all igneous rocks. This analysis is included in table 12.

If we want to compute the mineral composition corresponding to this average analysis, assumptions must be made as to the average abundance of undersaturated minerals (nepheline and olivine). Because Clarke did not include the extensive areas of the ocean floors in his average, it would seem that the amounts of olivine and nepheline must be rather small. We can follow Leith and Mead (1915) and put the amount of olivine as 2.6 per cent. Nepheline will be taken, quite arbitrarily, as 0.3 per cent.

In column I, table 13, the result of this computation is given. The composition of the alkali feldspar is supposed to be Or:Ab = 1:1.

Column II gives the average mineral composition of a mixture of intra-Pacific lavas and granite in the ratio 1:1. This corresponds to a mixture of the two dominant rock provinces in the earth. For the average granite the data given by Larsen (1942) were used.

Column III corresponds to the composition of the average igneous rock as computed by Leith and Mead (1915) by taking 65 per cent granite and 35 per cent basalt.

Column IV contains a "rough estimate" of the mean mineral composition of the igneous rocks as a result of a statistical examination of about seven hundred such rocks by Clarke and Washington (1924).

TABLE 13*

AVERAGE COMPOSITION OF IGNEOUS ROCKS

	I	II	III	IV
Quartz	12.4	14	20.4	12.0
Alkali feldspar	31.0	29	50.2	59.5
Plagioclase	29.2	28		
Pyroxene	12.0	13.5	12.9	16.8
Hornblende	1.7	0.5	1.6	
Biotite	3.8	2.5	3.9	3.8
Muscovite	1.4	(0.4†)	3.8	
Olivine	2.6	4.8	2.6	7.9
Nepheline	0.3	1.1	......	
Ores	4.1	5.2	4.6	
Chlorite and serpentine	0.6	(0.2†)	......	
Apatite	0.6	(0.6†)	......	
Titanite	0.3	(0.2†)	......	
Total	100.0	100.0	100.0	100.0

* Explanation of figures in columns given in text.

† Larsen gives no figures for these minerals in his average granite.

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OXIDATION AND REDUCTION IN GEOCHEMISTRY[1]

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ABSTRACT

Because many elements can occur in two or more oxidation states, reactions involving oxidation and reduction are important in geochemical processes. The energy change per equivalent of a reaction involving oxidation and reduction can be quantitatively expressed by its oxidation potential. For many inorganic reactions taking place in aqueous solution oxidation potentials are known and can be used to obtain a better understanding of geochemical processes. Conditions within the earth's crust may vary from highly oxidizing to highly reducing, and the oxidation potential of a natural environment determines what minerals may or may not be formed in that environment. The solution, transportation, and deposition of many elements in the earth's crust are strongly influenced by the oxidation potential of the environment. Separation of closely related elements, such as iron, cobalt, and nickel, may result. The significance of oxidation potentials in geochemistry is particularly marked in the frequent concentration and enrichment of rare elements in deposits formed under extreme oxidizing or reducing conditions.

INTRODUCTION

Many elements may occur in different oxidation states in the earth's crust. Commonest of such elements is iron, which occurs as the native metal (oxidation state 0), as ferrous compounds (oxidation state 2), and as ferric compounds (oxidation state 3). Similar elements are manganese (2, 3, 4), sulphur (−2, 0, 6), chromium (3, 6), vanadium (3, 4, 5), copper (0, 1, 2), cobalt (2, 3), lead (0, 2, 4), arsenic (0, 3, 5), antimony (0, 3, 5), silver (0, 1), nitrogen (−3, 0, 5), iodine (−1, 5), and many others. The concept of oxidation, which originally meant the addition of oxygen to a substance, now implies simply the increase in the oxidation-state number of an element, or, expressed in terms of the electronic theory for ionic compounds, the loss of electrons by an atom; reduction is, of course, the reverse process.

The stability of an element in a particular oxidation state depends upon the energy change involved in adding or removing electrons to transform it into a higher or lower oxidation state. Given a quantitative measure of this energy change for different reactions involving oxidations and reductions, it is then possible to arrange these reactions in order of intensity. Such a quantitative measure is provided by the factor known variously as the "oxidation-reduction potential," "oxidation potential," or "redox potential"; throughout this paper it will be referred to as the "oxidation potential." The concept of oxidation potential can be visualized in the following way: Consider a solution containing equal concentrations of the ions of the reduced and oxidized form of some element (e.g., ferrous and ferric iron), in which is immersed an unattackable electrode; a certain potential will be set up in that electrode, depending upon the tendency of the reduced form to give up electrons to the electrode and thereby transform to the oxidized form. The potential thus set up is the oxidation potential of the reaction by which the reduced form is transformed to the oxidized form and is signified by E_0. The oxidation potential of any reaction is a relative figure, the standard of reference being the reaction

$$H_2 = 2H^+ + 2e \quad (e = \text{electron})$$

(i.e., the removal of electrons from hydrogen atoms or the oxidation of hydro-

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gen to hydrogen ions). The oxidation potential of this reaction, for unit concentration[2] of the reacting substances (E_0), is arbitrarily fixed as 0.00 volt, and the scale of oxidation potentials extends on either side of zero.

Reactions involving oxidation and reduction can thus be rated according to intensity by their oxidation potentials. A positive value for an oxidation potential means that the oxidized form of the couple is a stronger oxidizing agent than hydrogen ions are and that the reduced form is a weaker reducing agent than is gaseous hydrogen; a negative value implies that the oxidized form is a weaker oxidizing agent than hydrogen ions are and that the reduced form is a stronger reducing agent than is gaseous hydrogen. The extreme values on the scale of oxidation potentials are for the following reactions:

$$2\text{HF} = \text{F}_2 + 2\text{H}^+ + 2e\ , E_0 = 3.03 \text{ volts}\ ;$$
$$\text{Cs} = \text{Cs}^+ + e\ , \qquad E_0 = -3.02 \text{ volts}\ .$$

In other words, gaseous fluorine is the strongest oxidizing agent known, and metallic cesium the strongest reducing agent. For positive values of E_0, reactions tend to go from right to left, for negative values of E_0, from left to right.

Table 1 lists a few of the many reactions whose oxidation potentials are known, in order of decreasing oxidation potential, i.e., in order of increasing reducing power, the reduced form of any couple having sufficient energy to reduce the oxidized form of any couple of higher positive potential.

The E_0 values are for the reactions at 25° C. when all the substances involved are at unit activity. The potential at other activities and temperatures is given by the expression

$$E = E_0 + \frac{RT}{nF} \ln Q$$

where Q is the product of the activities of the resultants divided by the product of the activities of the reactants, each activity raised to that power whose exponent is the coefficient of the substance in the reaction; ln signifies natural logarithms ($\ln x = 2.3026 \log x$); R is the gas constant (8.3151 volt coulombs); T is the absolute temperature; F is a faraday of electricity (96,500 coulombs); and n is the number of electrons involved in the reaction as written. This expression reduces to

$$E = E_0 + \frac{0.0002T}{n} \log Q$$

At 25° C. (298° absolute),

$$E = E_0 + \frac{0.06}{n} \log Q$$

The direct influence of temperature on oxidation potential is therefore very slight; for $n = 1$ and $\log Q = 1$ and with a change of temperature of 100° C., this would amount to only 0.02 volt. The indirect effect of temperature in changing the activities of the substances involved in the reaction would probably outweigh its direct influence. Variations in concentrations (activities) of the reacting substances do, however, have a marked effect on the oxidation potential of the reaction. For example, in the reaction

$$\text{Fe}^{2+} = \text{Fe}^{3+} + e\ , \quad E_0 = 0.77 \text{ volt}\ ,$$

if $\text{Fe}^{3+}/\text{Fe}^{2+}$ is 100/1, then E will be 0.89 volt; and if $\text{Fe}^{3+}/\text{Fe}^{2+}$ is 1/100, E will be 0.65 volt. This variation of oxidation potential with the concentration is of special importance in reactions involving hydrogen or hydroxyl ions, as many oxidation-reduction reactions do (table 1). In aqueous solutions, hydrogen-ion concentrations may vary from unity to 10^{-14} or, expressed in terms of pH, have pH from 0 to 14. Such a variation produces large changes in oxidation potentials involving hydrogen or hydroxyl ions and must be taken into account in applying E_0 values to actual reactions. The influence of pH on the oxidation potentials of some reactions given in table 1 is shown graphically in figure 1.

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[2] More correctly, "unit activity," activity being a function of concentration which provides for deviation from the laws of perfect solutions.

TABLE 1

	E_0 (in Volts)
$Co^{2+} = Co^{3+} + e$	1.84
$Ni^{2+} + 2H_2O = NiO_2 + 4H^+ + 2e$	1.75
$Au = Au^+ + e$	1.68
$Mn^{2+} = Mn^{3+} + e$	1.51
$Pb^{2+} + 2H_2O = PbO_2 + 4H^+ + 2e$	1.46
$Cr^{3+} + 4H_2O = HCrO_4^- + 7H^+ + 3e$	1.3
$Mn^{2+} + 2H_2O = MnO_2 + 4H^+ + 2e$	1.28
$2H_2O = O_2 + 4H^+ + 4e$	1.23
$VO^{2+} + 2H_2O = HVO_3 + 3H^+ + e$	1.1
$I^- + 3H_2O = IO_3^- + 6H^+ + 6e$	1.09
$VO^{2+} + 3H_2O = V(OH)_4^+ + 2H^+ + e$	1.00
$NH_4^+ + 3H_2O = NO_3^- + 10H^+ + 8e$	0.84
$Ag = Ag^+ + e$	0.80
$Fe^{2+} = Fe^{3+} + e$	0.77
$Ni(OH)_2 + 2OH^- = NiO_2 + 2H_2O + 2e$	0.49
$U^{4+} + 2H_2O = UO_2^{2+} + 4H^+ + 2e$	0.41
$4OH^- = O_2 + 2H_2O + 4e$	0.40
$Cu = Cu^{2+} + 2e$	0.34
$V^{3+} + H_2O = VO^{2+} + 2H^+ + e$	0.31
$I^- + 6OH^- = IO_3^- + 3H_2O + 6e$	0.26
$PbO + 2OH^- = PbO_2 + H_2O + 2e$	0.25
$Co(OH)_2 + OH^- = Co(OH)_3 + e$	0.2
$H_2S = S + 2H^+ + 2e$	0.14
$H_2 = 2H^+ + 2e$	0.00
$NH_3 + 9OH^- = NO_3^- + 6H_2O + 8e$	−0.12
$Cr(OH)_3 + 5OH^- = CrO_4^{2-} + 4H_2O + 3e$	−0.12
$Pb = Pb^{2+} + 2e$	−0.13
$Sn = Sn^{2+} + 2e$	−0.14
$Mn(OH)_2 + OH^- = Mn(OH)_3 + e$	−0.40
$Fe = Fe^{2+} + 2e$	−0.44
$S^{2-} = S + 2e$	−0.51
$Fe(OH)_2 + OH^- = Fe(OH)_3 + e$	−0.56
$Pb + 2OH^- = PbO + H_2O + 2e$	−0.58
$Zn = Zn^{2+} + 2e$	−0.76

The oxidation potential of a reaction gives information only about the state of chemical equilibrium and not about the rate at which this equilibrium is attained. The oxidation potential is a measure of the energy change involved in the reaction; but, even if the reaction is one which takes place with a decrease of free energy, it does not mean that the reaction will necessarily proceed at a significant rate under given conditions. For instance, the mechanism of the reaction may be such that it cannot take place in spite of a favorable potential value. Thus $Fe = Fe^{3+} + 3e$, $E_0 = -0.04$ volt indicates that H^+ should oxidize iron to the ferric state; but the mechanism of the reaction is $Fe = Fe^{2+} + 2e$, $E_0 = -0.44$ volt, and $Fe^{2+} = Fe^{3+} + e$, $E_0 = 0.77$ volt and H^+ cannot bring about the second step. In general, it may be stated that many reactions for which oxidation potentials have been calculated are not capable of experimental attainment under equilibrium conditions for one of two reasons: (*a*) the rate of the reaction is exceedingly slow, or (*b*) the final and initial substances cannot exist together because of an intermediate state, as, for example, the Fe–Fe^{3+} couple instanced above.

THE OXIDATION POTENTIAL OF NATURAL ENVIRONMENTS

Before proceeding to the discussion of the application of oxidation potentials to specific geological and geochemical problems, it is desirable to consider in some detail the factors controlling the intensity of oxidizing and reducing processes which take place in nature. Brief consideration will make clear the fact that many reactions for which oxidation potentials are given in the literature never take place in natural environments (for example, the oxidation of sulphate to persulphate or the reduction of titanium compounds to metallic titanium). This indicates that the range of oxidation potentials of natural environments is limited, thereby limiting the reactions which may or may not take place.

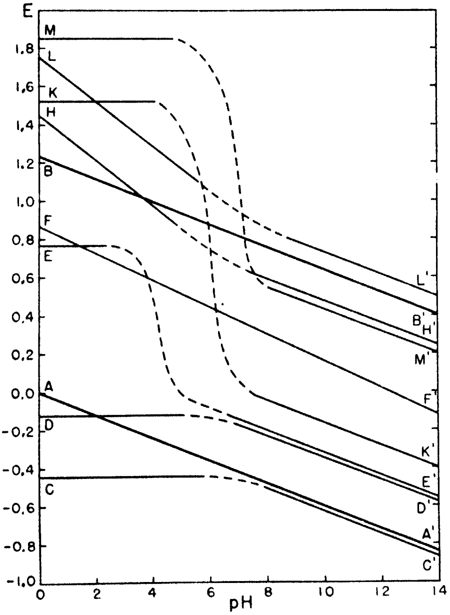

FIG. 1.—Variation of oxidation potential with pH for certain reactions

A. $H_2 = 2H^+ + 2e$
B. $2H_2O = O_2 + 4H^+ + 4e$
C. $Fe = Fe^{2+} + 2e$
D. $Pb = Pb^{2+} + 2e$
E. $Fe^{2+} = Fe^{3+} + e$
F. $NH_4^+ + 3H_2O = NO_3^- + 10H^+ + 8e$
H. $Pb^{2+} + 2H_2O = PbO_2 + 4H^+ + 2e$
K. $Mn^{2+} = Mn^{3+} + e$
L. $Ni^{2+} + 2H_2O = NiO_2 + 4H^+ + 2e$
M. $Co^{2+} = Co^{3+} + e$

A'. $H_2 + 2OH^- = 2H_2O + 2e$
B'. $4OH^- = O_2 + 2H_2O + 4e$
C'. $Fe + 2OH^- = Fe(OH)_2 + 2e$
D'. $Pb + 2OH^- = PbO + H_2O + 2e$
E'. $Fe(OH)_2 + OH^- = Fe(OH)_3 + e$
F'. $NH_3 + 9OH^- = NO_3^- + 6H_2O + 8e$
H'. $PbO + 2OH^- = PbO_2 + H_2O + 2e$
K'. $Mn(OH)_2 + OH^- = Mn(OH)_3 + e$
L'. $Ni(OH)_2 + 2OH^- = NiO_2 + 2H_2O + 2e$
M'. $Co(OH)_2 + OH^- = Co(OH)_3 + e$

This limitation is due to the fact that chemical reactions at or near the earth's surface take place, in general, in aqueous solution. Theoretically, this limits the reactions which can occur to those with oxidation potentials lying between those for the reactions

$$H_2O = \tfrac{1}{2}O_2 + 2H^+ + 2\,e\,,\ E_0 = 1.23\text{ volts}\,,$$

and

$$2H^+ + 2\,e = H_2\,,\qquad E_0 = 0.00\text{ volt}\,.$$

The oxidized form of any couple with a higher potential than that for the first reaction will theoretically decompose water with the evolution of oxygen and hence cannot be formed in aqueous solution. The reduced form of any couple with a lower potential than that for the second reaction will theoretically decompose water with the evolution of hydrogen and similarly cannot be formed in aqueous solution. It is found that the theoretical requirements are not strictly met in practice, on account of overvoltage phenomena, i.e., it requires a greater potential than the theoretical to produce the evolution of hydrogen or oxygen at a measurable rate. However, a study of the reactions actually taking place in nature indicates that the potentials of these two reactions do largely control the nature of the oxidations and reductions that take place under natural conditions.

Both these reactions involve hydrogen ions, and their potentials are thus strongly affected by changes in hydrogen-ion concentration. The E_0 values given above are, of course, for hydrogen-ion concentrations of unity, i.e., pH = 0, and the potentials decrease (at 25° C.) 0.06 volt for each tenfold decrease in hydrogen-ion concentration (each unit increase in pH). The pH of natural waters is very variable, ranging from as low as 0 in strongly acid waters of volcanic regions to 10 or more in alkaline areas where sodium carbonate is present in solution. As a rule, however, the pH of natural waters lies between 4 and 9, the great majority being within one unit of the figure for pure water (pH = 7). For a pH of 7 the potential of the first reaction would be 0.82 volt and of the second reaction would be −0.41. This gives a range of 1.23 volts, compared to the total range of the scale of oxidation potentials of about 6 volts; therefore but a comparatively small number of oxidation-reduction reactions may be expected to take place under natural conditions.

The above figures indicate that the oxidation potentials of natural environments, where the pH is near 7, should lie between −0.41 and 0.82 volts. Unfortunately, comparatively few measurements of the oxidation potentials of natural environments have been made. ZoBell (1946) has made a number of measurements of oxidation potentials in recent marine sediments on the California coast and obtained values ranging from 0.350 (in sandy deposits from shallow, well-oxygenated water) to −0.500 volt (in fine-grained bottom deposits rich in organic matter), the pH ranging from 6.4 to 9.5. Allgeier, Hafford, and Juday (1941) found the oxidation potentials of surface waters in Wisconsin lakes to be 0.38–0.50 volt, of bottom water 0.057–0.44, and of bottom sediments, −0.14–0.2. Hutchinson, Deevey, and Wollack (1939), also investigating lake waters, found potentials of 0.4–0.5 volt in surface waters, and potentials as low as 0.10 in bottom waters. These rather meager data in general fall well within the limits predicted by the foregoing theoretical discussion. It will be noted, however, that in fine-grained bottom deposits rich in organic matter ZoBell found potentials as low as −0.500 volt, more negative than the theoretical potential of −0.42 volt required to liberate hydrogen from water at pH 7. ZoBell ascribes this creation of a hydrogen overvoltage to the presence of anaerobic bacteria, which thrive in such environments.

It would be interesting to have more measurements of the oxidation potentials of natural waters, particularly from more unusual environments, such as desert areas, where conditions are extremely oxidizing, and from mineral springs and mine waters. Because these measurements can readily be made with the same equipment as that used extensively for the electrometric measurement of pH, it is not too much to hope that, where measurements of pH of natural waters are being made, measurements of oxidation potential will be made at the same time.

A study of mineral associations enables predictions to be made regarding the approximate oxidation potential under which the minerals were formed. The rare occurrence of metallic iron in carbonaceous sediments indicates an extremely low oxidation potential at the time of formation. Another strongly reducing environment must have been that which gave rise to the association of native lead and pyrochroite [$Mn(OH)_2$] found in veins in the mine at Långban, Sweden. Ferrous compounds are generally considered as indicative of strongly reducing conditions, but reference to figure 1 shows that, in strongly acid solutions, ferrous iron is not readily oxidized to the ferric state. At the other end of the scale the occurence of minerals containing chromate, nitrate, or iodate or of plattnerite (PbO_2) or stainierite (CoOOH) indicates strongly oxidizing conditions of origin.

APPLICATION OF OXIDATION POTENTIALS TO GEOCHEMICAL PROBLEMS

The concept of oxidation potentials is capable of wide application in geochemistry, as the solution, transportation, and deposition of those elements which may occur in two or more oxidation states are directly and powerfully influenced by the oxidation potential of the environment. The application of oxidation potentials to geochemical problems was first discussed by Goldschmidt (1933) and more recently by Scerbina (1939), Mason (1941), and Chapman and Schweitzer (1947).

THE OCCURRENCE OF NATIVE ELEMENTS

The occurrence or nonoccurrence of elements in the native state is readily explained on the basis of oxidation potentials. The higher the potential required to remove electrons from a metal and convert it into metallic ions, the greater the tendency for the metal to be found in the native state. On the other hand, if an element has an oxidation potential lower than that of the hydrogen electrode, it is theoretically impossible to discharge its ions in aqueous solution, so that it can be formed only in a nonaqueous environment. Observations on the occurrence of native metals bear out these contentions. Gold and platinum, which have oxidation potentials greater than 1 volt, occur almost entirely as native metals. Copper, silver, and mercury, which have oxidation potentials between 0 and 1 volt, are found both as native metals and in compounds. Metals with negative oxidation potentials occur native very rarely and have evidently been formed under special circumstances. Native lead has been found at a number of localities; its oxidation potential is only slightly negative (−0.13 volt), and reference to figure 1 shows that in alkaline and weakly acid solutions its oxidation potential is greater than that of the hydrogen electrode and therefore it could be deposited from such solutions. Tin, with an oxidation potential of −0.14 volt, has also been reported in the native state. Most of the terrestrial occurrences of native iron are in igneous rocks; but it has been found in carbonaceous sediments. The negative oxidation potential of −0.44 volt for iron would at first seem to preclude the deposition of native iron from aqueous solutions; but reference to figure 1 reveals that, on account of the rapid decrease in the potential of the hydrogen electrode with increase of pH and of the overvoltage required to discharge hydrogen ions, iron could be deposited from solutions of low hydrogen-ion concentration; the oxidation potential of −0.500 volt recorded by ZoBell (1946) for a marine mud indicates that under suitable conditions metallic iron could be produced in such an environment. The only element with oxidation potential lower than that of iron which has been recorded as occurring in the native state is zinc (−0.76 volt). All records of native zinc are extremely dubious, and the reported native zinc is probably of artificial origin. Of the many elements with oxidation potentials less than that of zinc—the alkali and alkaline-earth metals, aluminum, titanium, chromium, manganese—none is known to occur native.

STEPWISE OXIDATION OF IRON-MANGANESE MINERALS

The oxidation of minerals containing both iron and manganese affords an interesting example of the influence of oxidation potentials on mineral formation. There are a number of phosphate minerals containing both divalent iron and manganese—triphyllite, lithiophilite, natrophilite, arrojadite, and graftonite; they alter readily and are remarkable for the number and diversity of their alteration products. They all show a typical stepwise alteration, characterized by the complete oxidation of the iron to the trivalent state before the divalent manganese is affected. This alteration gives rise to a number of mineral species—sicklerite, alluaudite, landesite, salmonsite, etc.—containing divalent manganese and trivalent iron. The oxidation potential required to convert ferrous to ferric iron is so much lower than that required to convert divalent manganese to the trivalent or tetravalent state that the oxidation of the ferrous iron proceeds to completion before the manganese is affected.

Incidentally, it is interesting to note that a much higher potential (1.51 volts) is required to convert manganous ions into manganic ions than is required to convert manganous ions into manganese dioxide (1.28 volts). Furthermore, the latter potential decreases rapidly with increase in pH, whereas the former is independent of pH, at least in solutions sufficiently acid to prevent the precipitation of hydroxides. This accounts for the rarity of compounds containing trivalent manganese as minerals; under normal conditions oxidation of manganous compounds results in the formation of pyrolusite and psilomelane. The formation of purpurite [$(Mn,Fe)PO_4$] by the oxidation of the phosphate minerals mentioned above suggests that the presence of phosphate ion greatly reduces the normal potential required to oxidize manganous to manganic ions, probably on account of complex formation between manganic and phosphate ions.

CONDITIONS OF FORMATION AND STABILITY OF MINERALS

The oxidation potential of the environment is an important factor in the formation of minerals, and therefore from the nature of a mineral association it is often possible to ascertain the conditions of paragenesis. Reverting to the occurrence of native lead at Långban in Sweden mentioned previously, the native lead occurs associated with pyrochroite [$Mn(OH)_2$] in veins which were evidently formed at comparatively low temperatures from aqueous solutions. Study of figure 1 shows that the oxidation potential of the medium must have been negative to permit deposition of metallic lead and that deposition probably took place in an alkaline or only weakly acid solution, because in strongly acid solution the oxidation potential of lead is less than that of the hydrogen electrode. The alkaline nature of the medium is confirmed by the occurrence of pyrochroite, as the precipitation of $Mn(OH)_2$ requires a pH of 8 or greater.

Similarly, the high potentials required to convert divalent cobalt to trivalent cobalt and divalent lead to PbO_2 in acid solution indicate that the minerals stainierite (CoOOH) and plattnerite (PbO_2) are deposited from alkaline solutions, for which the oxidation potentials are much less (fig. 1). Figure 1 shows that for many reactions the oxidation potentials decrease rapidly with the increase of pH, generally more rapidly than does the corresponding decrease in the oxygen and hydrogen electrodes. Thus oxidation in general proceeds more readily, the more alkaline the solution. The potentials for some oxidations in alkaline solutions, particularly those which result in the precipitation of practically insoluble compounds, lie far below the potentials for corresponding oxidations in acid solutions. This is especially marked with respect to the oxidation of ferrous to ferric iron; in acid solution the potential is 0.77 volt and is not affected by pH; however, as soon as the pH increases to a figure at which ferric hydroxide is precipitated (about pH 2–3) the oxidation potential drops sharply to a negative figure. Thus ferrous salts are comparatively stable in acid solution, being only slowly oxidized by air; but in solutions sufficiently low in acid for $Fe(OH)_3$ to be precipitated, oxidation proceeds rapidly to completion. Deposition of ferrous compounds in nature therefore demands either a very acid environment or one with a very low oxidation potential, on the negative side of zero.

SEPARATION OF ELEMENTS ON THE BASIS OF OXIDATION POTENTIALS

The separation of closely related elements in the upper zone of the lithosphere by processes involving solution and redeposition is often brought about by their distinctive properties with respect to oxidation and reduction.

A very good example is that of the three elements iron, nickel, and cobalt. Generally they occur together in primary deposits, yet supergene processes result in their separation. Figure 1 shows that these three elements differ greatly in the potentials required to oxidize them beyond the divalent state. Iron is readily oxidized to the trivalent state in alkaline and mildly acid environments; cobalt requires a much higher potential even in alkaline solution, and in acid solutions the potential required lies high above that for the oxygen electrode; nickel does not form trivalent compounds, but a dioxide is known, the formation of which, even in alkaline solutions, requires potentials somewhat higher than that for the oxygen electrode. This is reflected in natural occurrences: the common form of iron in supergene deposits is as hydrated ferric oxide; hydrated cobaltic oxide (stainierite) is found only where conditions have been strongly oxidizing; and the higher oxide of nickel is not known to occur as a mineral. The separation of these three elements by supergene processes is well illustrated in New Caledonia, where serpentinized ultra-basics altering under tropical conditions give rise to lateritic material rich in Fe_2O_3 and concentration of nickel as garnierite and of the cobalt as pockets of hydrated cobaltic oxide. In this connection the work of Leith and Mead (1916) on the lateritic iron ores of eastern Cuba is pertinent. On the basis of chemical analyses of the residual iron ore and of the serpentine from which it was derived, they showed that, in contrast to the enrichment in iron in the residual deposit, the nickel and the cobalt are strongly decreased in absolute amount and even more strongly in relative amount; these two elements have evidently been largely removed in solution on account of their greater stability in the divalent state.

Oxidation processes also result in a similar separation of manganese and iron. Manganese is often closely associated with iron in primary deposits, the ferrous and the manganous ions being similar in their chemical nature. Supergene processes, however, generally lead to a complete separation of iron from manganese; because the potential required to convert iron to the ferric state is so much lower than that required to convert manganese to manganese dioxide, the iron is readily precipitated as hydrated ferric oxide, whereas the manganese remains in solution and is eventually deposited under more oxidizing conditions as comparatively iron-free manganese dioxide.

ENRICHMENT OF RARE ELEMENTS IN OXIDIZING OR REDUCING ENVIRONMENTS

It has been observed that rare elements are often concentrated by natural processes in unusual environments. These environments are often characterized by very high or very low oxidation potentials. A good example is the marked concentration of iodine in the highly oxidizing environment of the nitrate deposits of northern Chile, the iodine having been fixed in the form of iodates. The ores of copper, lead, vanadium, and uranium which are found in sandstones and shales far from igneous rocks, generally represent deposits formed under desert conditions in oxidizing environments; these deposits often show marked concentrations of rarer elements also, especially silver, nickel, cobalt, chromium, molybdenum, and selenium. Analyses of natural deposits of manganese dioxide, in the form of pyrolusite and psilomelane, frequently show notable amounts of rarer elements, especially cobalt, copper, and molybdenum, sometimes also lead and tungsten. At the other end of the scale, in highly reducing environments, rarer elements are also enriched. Such reducing environments are the present-day black marine muds rich in organic matter and anaerobic bacteria and, on land, areas of accumulation of decaying vegetable matter, such as peat swamps. In the geological column these are represented by bituminous and carbonaceous shales. Goldschmidt (1935) found in coal ash marked concentration of cobalt, nickel, gallium, germanium, molybdenum, and arsenic. In this connection the frequent association of millerite (NiS) with coal is noteworthy. Similarly, the ash of natural hydrocarbons generally shows considerable amounts of vanadium and often also of nickel and molybdenum. The Mansfeld copper slate in Germany, probably a "fossil" black marine mud, is an example of a bituminous shale which is worked for its copper content and which also shows notable concentration of vanadium, arsenic, antimony, molybdenum, cadmium, silver, gold, and metals of the platinum group (Cissarz and Moritz, 1933). It is not clear to what extent the original concentration of rare elements in these reducing environments took place as a result of the vital activities of plants and animals or during subsequent processes of decay and mineralization, but in any case the fixation of these rare elements was largely controlled by the strongly reducing conditions prevailing in these sediments.

Such concentrations of rare elements suggest that sedimentary deposits formed under strongly oxidizing or reducing conditions deserve careful examination for unexpected concentrations of rare elements, which under favorable circumstances may be present in workable amount, either alone or as a by-product of the exploitation of some common ore. In this connection sedimentary manganese deposits represent a commonly occurring environment of high oxidation potential, and carbonaceous and bituminous shales an environment of low oxidation potential. Such materials may eventually prove to be important sources of these rare elements whose industrial importance is continually growing.

THE OXIDATION POTENTIAL OF NONAQUEOUS ENVIRONMENTS

The theory and applications of oxidation potentials have been developed from a consideration of reactions taking place in aqueous solutions, and practically no progress has yet been made toward applying the theory to reactions taking place in nonaqueous environments. Because most reactions taking place in the upper zone of the lithosphere undoubtedly are "wet," the utility of this theoretical background in a consideration of these reactions is very great. Nevertheless, the broad field of magmatic reactions cannot be attacked satisfactorily at present because oxidation potentials determined in aqueous solutions at room temperatures can hardly be extrapolated to the high temperatures and otherwise very different conditions of a silicate melt. Yet reactions involving oxidation and reduction undoubtedly take place in magmas. All magmas contain iron and hydrogen and oxygen in some form, and reactions involving equilibria between divalent and trivalent iron and between hydrogen, oxygen, and water must play a large part in determining the minerals which are produced on crystallization. In general, the magmatic environment is a reducing environment, as in minerals formed by direct magmatic crystallization the elements are generally present in the lowest oxidation state in which they occur in nature. In plutonic rocks, especially the more basic, ferrous iron is commonly predominant over ferric iron, and even in volcanic rocks, which have had greater opportunity to acquire oxygen, the same relation generally holds. Titanium may be present in igneous rocks to some extent in the trivalent state; the violet color and pleochroism of titanium-bearing augite is suggestive of the Ti^{3+} ion. Incidentally, the most reducing environment represented in nature is probably that of meteorites, which contain metallic iron, oldhamite (CaS), and daubreelite (Cr_2FeS_4). There is undoubtedly a wide and interesting field in the investigation of oxidation and reduction in a magmatic environment, but experimental difficulties are great and the theory largely undeveloped.

OXIDATION AND REDUCTION IN THE GEOCHEMICAL CYCLE

The geochemical cycle in the lithosphere can be conceived as proceeding from the initial crystallization of a magma through the alteration and weathering of the igneous rock; the transportation and deposition of the material thus produced; then the cycle continues through diagenesis and lithification to metamorphism of successively higher grade until eventually, by processes of anatexis and palingenesis, a magma is regenerated. Like any ideal cycle, the geochemical cycle may not be realized in practice; the cycle may be indefinitely halted at some stage, or its direction may be reversed; but nevertheless it provides a useful concept to serve as a basis for discussion of many aspects of geochemistry.

The processes of oxidation and reduction are distinctly correlated with different stages in the geochemical cycle outlined above. As previously pointed out, the magmatic environment is a strongly reducing one, and rocks formed by direct magmatic consolidation generally show a low oxidation state. Alteration and weathering take place at or near the earth's surface, where the presence of free oxygen produces a high oxidation potential; and in this part of the cycle the highest oxidation state is reached. Diagenesis and lithification may result in some degree of reduction, especially in sediments rich in organic matter; and metamorphism, especially in its higher grades, quite generally results in progressive reduction until finally the material is brought back to its originally magmatic condition. Nevertheless, the imprint produced on the material by oxidation at the earth's surface may persist throughout the metamorphic part of the cycle, and often a high state of oxidation of a metamorphic rock (which may be most readily recognized by a high Fe_2O_3/FeO ratio) is a useful indication of the original sedimentary character of the rock.

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CELLULAR STRUCTURES IN GLASS AS RELATED TO STRUCTURES IN LAVA[1]

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ABSTRACT

Cellular structures which form during the cooling of some unusual optical glasses are described. These resemble in several respects ellipsoidal structures in lava. The former are shaped by convection currents present in the cooling liquid and remain visible in the chilled material because of chemical and physical differentiation of the liquid. It is suggested that some of the structures in lava variously called ellipsoidal, pillow, or globular may have formed by convection flow modified by horizontal movement of the lava stream and followed by rather rapid quenching.

GENERAL STATEMENT

The origin of structures which may develop in a body while changing from a fluid to a solid has long been an interesting subject of discussion (Bénard, 1900; Dauzère, 1908; Lewis, 1914; Sosman, 1916; Foye, 1924; Fuller, 1931; and Stark, 1938). The structures occurring commonly in basalt, called ellipsoidal, pillow, or globular, are generally assumed to have formed during extrusion into water, the rapid chilling of the lava being a condition of their formation. Columnar or prismatic structures in lavas no doubt usually result from contraction of the layer during cooling, but formation of columnar structures chiefly as a result of convection flow of the fluid lava during cooling is a possible explanation for some occurrences of this structure (Longchambon, 1913; Sosman, 1916, p. 223).

The writer recently had the opportunity of studying structures which form during the cooling of some optical glasses.[2] These structures resemble in some respects those occurring in lavas; and inasmuch as there is little doubt as to the general mechanism of their formation, it has seemed worth while to describe them.

DESCRIPTION OF STRUCTURES

Examples of cellular structures in glass are shown in plate 1 and figures 1 and 2. In plate 1, *C*, is shown a typical specimen exhibiting cellular structure. The composition of the mixture was such that, on cooling, the glass devitrified sufficiently to make it opaque except for thin layers separating the cells. The cellular structure developed while the glass

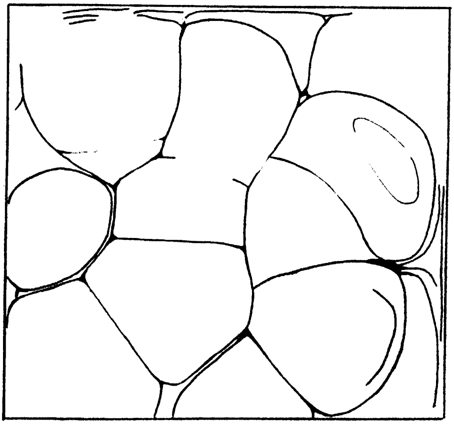

FIG. 1.—Specimen of glass having cellular structure. This specimen devitrified on cooling except for cell borders, which remained a glass. Area shown is approximately 6 inches across.

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[1] Manuscript received March 10, 1948.

[2] Experimental glasses made at the Eastman Kodak Co., Rochester, N.Y.

was fluid and before devitrification occurred, but the subsequent devitrification, which did not affect the glass septa at the borders of the cells, has made the structure easily visible. The glass was poured into a square, flat-bottomed mold, flowing across the mold in a direction from bottom to top in this photograph. The ridge of glass in the central part of the bottom edge joined the lip of the crucible before the latter was removed. There is a tendency for the cells to be arranged in concentric rows about this source position.

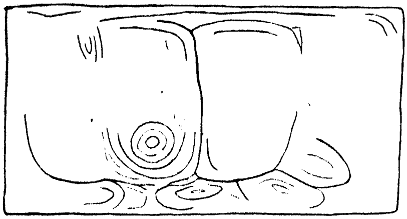

FIG. 2.—Specimen of glass having cellular structure. This specimen partly devitrified during heat treatment subsequent to cooling. Cells are concentrically zoned. Short dimension approximately 3 inches.

A view of the top surface of three other blocks of glass exhibiting cellular structure is shown in plate 1, *B*, and figures 1 and 2. In figure 1, septa of glass delimit the cells, which devitrified during cooling. The specimen in figure 2 was entirely glass when cooled to room temperature in its mold. The devitrification, which occurred selectively within the cells, leaving the borders glassy and thereby emphasizing the cellular structure, was produced by a later heat treatment. A concentric zoning within the cells can be faintly seen. Such zoning is commonly visible within cells on close inspection but is not usually a pronounced feature. The specimen in plate 1, *B*, is all glass except for an accumulation of crystallites on the top surface. These crystallites tended to accumulate at cell intersections, thereby outlining the cells which otherwise would not be discernible in the photograph.

In plate 1, *A*, is shown the cross section and top of a specimen devitrified except for the glass border which occurs along the bottom and top as well as the sides of the cell. Commonly, the glass septa do not completely close under the cells, however, resulting in a short, columnar type of structure.

The border of the cells, as well as the top surface to the same order of thickness as the border, has a slightly different composition from the bulk of the cell as indicated by differences in refractive index. In one specimen which did not devitrify, the index of the main body of the glass and of the cell borders is, respectively, 1.65 and 1.75.

The top surface of the glass has a microscopically cellular structure. In the central part of a large cell surface (pl. 2, *A*) the tiny polygons are roughly equidimensional. Toward the borders of the large cells (pl. 2, *B*) the small polygons

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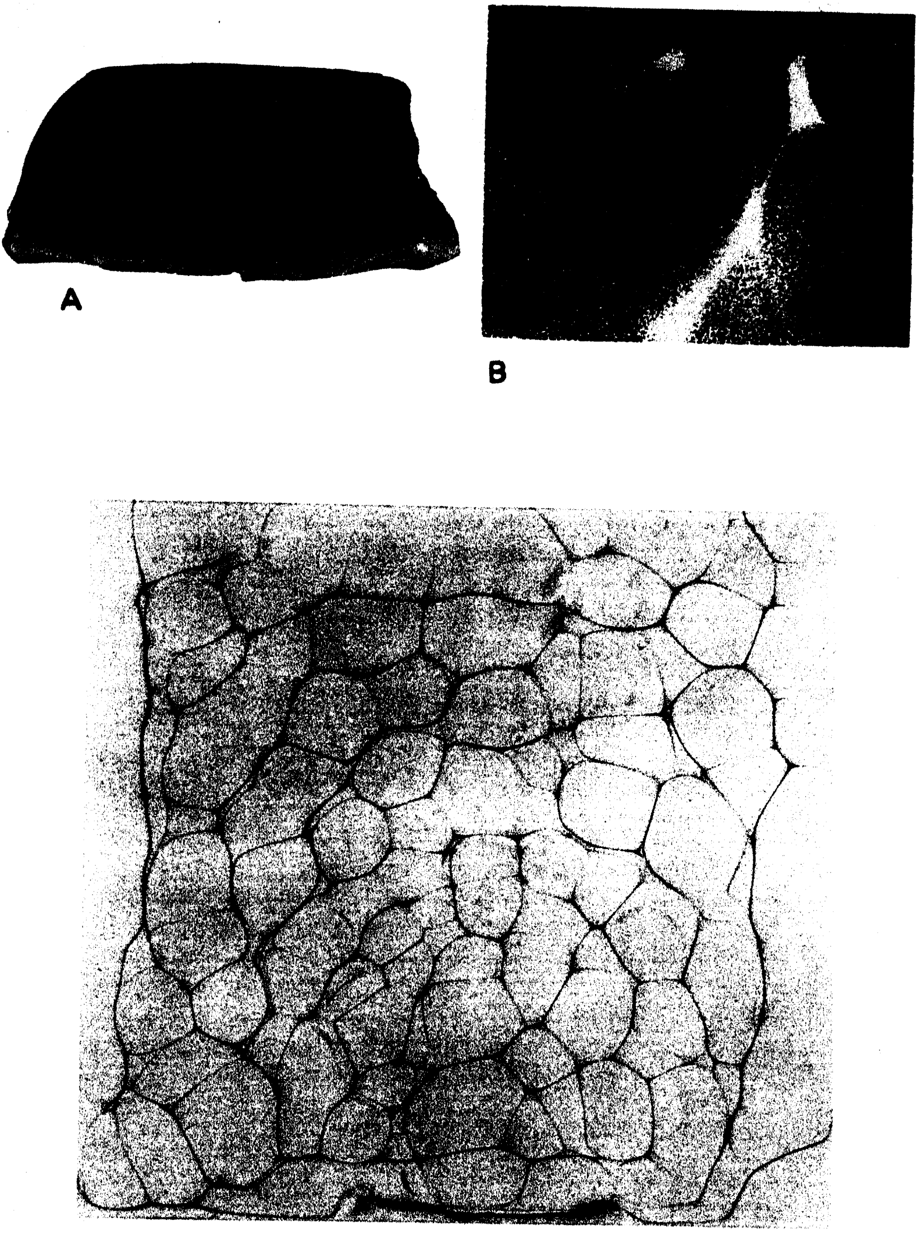
PLATE 1

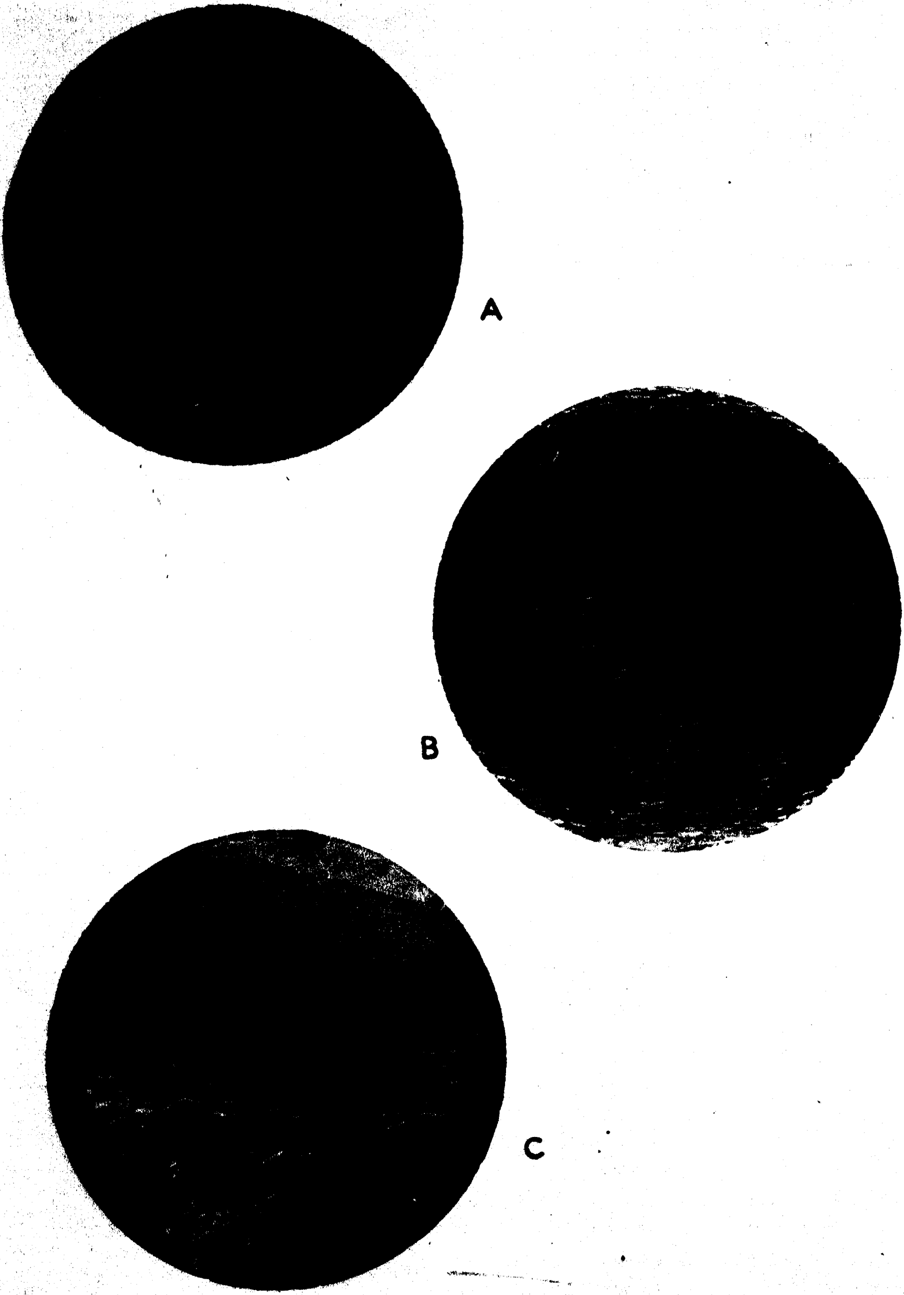
A, Top and side view of a glass specimen approximately 2 inches long, showing cross section of a cell.

B, Top surface of a specimen of glass having cellular structure. Some cells are outlined by accumulation of surface crystallites.

C, Top surface of a glass specimen exhibiting cellular structure. Specimen is 10 inches across and ⅜ inch thick. Along the bottom edge of the specimen is a ridge of glass, 4 inches long, which has the position of the lip of the platinum crucible during pouring. Note tendency toward concentric arrangement of rows of cells about this source position.



Cellular structures in glass



Cellular structures in glass

become elongated and the junctions of the large cells have a schistose appearance. Plate 2, *C*, is a photomicrograph of a part of the surface of the specimen shown in plate 1, *B*. Crystallites are present in this border zone, which grades into the central zone of regular polyhedra toward the top of the photograph.

DESCRIPTION OF THE GLASS

The glass in which the structures described above developed is a fluosilicate, composed of about equal amounts by weight of NaF, TiO_2, and SiO_2. These fluosilicate glasses are more fluid and more fuming than the usual commercial glasses owing to the large amount of fluorine substituting for oxygen in the structure. The glasses are unusual and of interest to the optical designer because of their high dispersion for a given value of n_D, but for optical use striae must be eliminated from the glass, and therefore cellular structures are obviously undesirable. Liquids in the system NaF–TiO_2–SiO_2 readily devitrify on cooling except for a limited series of compositions. The specimens shown in plate 1 and figures 1 and 2 are on the borders of the glass field in this system.

These experimental glasses were made usually in 3,000-gm. melts. The batch is melted and stirred in a platinum crucible until homogeneous, following a technique developed at the Eastman Kodak Company Research Laboratories for manufacture of the rare-earth glasses. When homogeneous and at a temperature slightly above its liquidus, the fluid glass is poured rapidly into a flat-bottomed, cast-iron mold and chilled. A slab of glass about 1 inch thick and 6–8 inches across is the usual result.

ORIGIN OF THE STRUCTURES

The cellular structures here described are a consequence of convection currents set up in the upper half or two-thirds of the glass slab during chilling. Heat is rapidly removed from the lower part of the glass layer by the cast-iron bottom of the mold. Convection currents do not tend to form in the lower part of the glass, therefore, because the temperature of the liquid increases upward. But loss of heat and volatiles from the top of the liquid produces an inverted density gradient in the upper part of the layer of glass which is sufficient to set up convection currents in the fluid glass, and these persist until the viscosity is too great to permit movement of the liquid. The convection currents divide the upper layer of the glass into cells, their diameter being dependent on factors such as thickness of layer, viscosity, and temperature gradient. Within each cell the cooled, more dense surface layer flows to the border; moves down the side to a warmer region; passes along the bottom of the cell; and then, having expanded, completes the circulation by moving upward in the central part of the cell. Such currents, if viscosity and temperature gradient remained constant until a steady state of flow developed, would produce a columnar struc-

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PLATE 2

A, Photomicrograph of top surface of glass in central part of a cell. Area shown is about 5 mm. across.

B, Photomicrograph of junction of two cells, top surface of glass. The dark band dividing area into about equal parts is glass border of cells. Scale same as in *A*.

C, Photomicrograph of a portion of top surface of specimen in plate 1, *B*, showing accumulation of crystallites along border of cells. Area shown in approximately 1½ mm. across.

ture composed of vertical, six-sided prisms. But in these glasses viscosity and temperature gradient change rapidly, the liquid mass has a horizontal and turbulent flow for the first few seconds as it fills the mold, and the convection flow is halted shortly after it begins by the rapid increase in viscosity of the glass. Consequently, well-developed columnar structure rarely appears. Globular shapes with flat tops and frequently rather indistinct bottoms, and forms intermediate between this and columns, are usual.

Convection cells are visible in the hot liquid almost immediately on its being poured and sometimes can be seen in the liquid as it is flowing across the mold. The cell outlines appear as red lines of a darker color than the central part of the cells. During the first few seconds after the mold is filled the cell borders shift slightly in position as thermal gradient and viscosity change, but the final pattern of cells is essentially the same as that which first appears, and commonly this pattern is affected by the direction and speed of movement of the liquid across the mold. In plate 1, *C*, a tendency toward a concentric arrangement of rows of cells about the point of issuance of the liquid into the mold can be seen.

Cellular structures are preserved in these fluosilicate glasses because of the properties of differentiated liquids and rarely by accumulation of crystals. The borders of the cells are a differentiate of the glass both chemically and physically. The top layer, which by flowage becomes the border, loses volatiles and cools as it moves from the central part of a cell to the border. The loss of volatiles results in the glass having a higher refractive index. Both the loss of volatiles and the more rapid cooling make the border glass more resistant to devitrification than that in the interior of a cell. If crystallites form in the surface layer, these may be carried by the glass currents to the cell borders outlining the cells, a process which apparently occurred during cooling of the specimen shown in plate 1, *B*.

RELATION OF STRUCTURES TO THOSE IN LAVAS

A fluid layer of basaltic lava flowing over a surface of low relief and cooled principally by heat loss from its upper surface should contain convection currents. Inasmuch as the rock over which the flow spreads is a poor conductor and might even be a previous warm flow, heat removal from the base would be relatively slow in contrast to the rapid removal of heat from the base of a fluosilicate glass layer by cast iron. Convection flow therefore would be expected to affect nearly the entire thickness of the lava layer, causing a chilled surface layer to move downward toward the bottom along a curved path. Horizontal movement of the lava coupled with convection flow might produce ellipsoidal structures visible because of chilled borders. These structures could be preserved if the cooling of the lava, as by contact with water, were sufficiently rapid. Were the lava to come to rest long enough for an approach to a steady state of convection flow to appear, resulting columnar structures might be visible in the solidified rock.

If the lava were extruded over a wet surface, absorption of an appreciable amount of water by the silicate liquid might occur, as described by Sosman for the case of blast-furnace slags (Sosman, 1947, p. 287). The higher water content of the lower part of the layer of lava would lower both its density and its viscosity, and this could be an additional factor in producing convection flow in lava streams.

Experiments with fluosilicate glasses

are on such a small scale as compared with lava flows that comparisons may not be justified. Some field observations, however, are suggestive that ellipsoidal structures in lavas may result from a combination of convection and lateral flow. In describing the great outpouring of lava at Mauna Loa in 1859, Green states (1887, p. 163):

> When it first broke out the lava formed a fountain of white-hot highly liquid lava, which played steadily for many days, gradually lowering, however, till finally it simply ran over, flowing down the mountain under its own cooled crust without any sign of steam, vapor or gases proceeding from it from its source to the sea at Wainanalii, a distance of about forty miles. Here we watched this molten lava pouring into the sea for several hours. It ran over a low shelf about ten feet high, and extended perhaps 500 or 600 feet wide, and fell into the sea where it was about 20 or 30 feet deep. It came from under the crust in great red-hot flattened spheroidal masses, having something the appearance of masses of moderately thick porridge as it is poured from a saucepan. The spheroidal masses, however, being perhaps 10 feet to 15 feet wide, and 4 to 6 feet deep. There was no steam, vapor or gas whatever to be seen coming from this lava till it went underwater. Indeed the first contact of the red-hot spheroids did not seem to produce a particle of steam, and it was only when each had gone under water and become partially cooled off, that a puff of steam rose above the surface.[3]

Green's "great red-hot flattened spheroidal masses" of which the lava layer was composed are considered by the writer to be convection cells, similar but on a larger scale to those in a fluosilicate liquid as it moves across the mold. Although the spheroidal masses in the lava are much larger, the ratio of width to depth is about the same as is usual for cells in the fluosilicate glass. It is considered probable by the writer that on being immersed in the water the spheroidal masses maintained their identity and built up a deposit of pillow lava.

In the subglacial extrusions of basalt in Iceland, globular basalt similar to pillow lavas of other areas is common (Noe-Nygaard, 1940). In describing one occurrence Noe-Nygaard states (1940, p. 57): "Following the globular basalt in a southerly direction we find that it passes smoothly into a basalt without globular structure, but with pronounced columnar structure (irregularly oriented columns)." The smooth gradation of globular into columnar structure suggests that the origin of both was associated with convection flow, a more nearly steady state of flow having been reached where a columnar structure developed.

Acknowledgments.—The writer wishes to thank Mr. P. F. DePaolis of the Eastman Kodak Company for his co-operation in supplying most of the glass specimens used in this study. He is indebted also to Dr. H. E. Merwin, Dr. R. B. Sosman, and Professor J. T. Stark for their criticisms of the manuscript.

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[3] Some additional information about this lava flow is given in Hitchcock (1911), p. 103. Hitchcock states: "Mr. W. L. Green observed the entrance of the lava into the sea, both in January and several months later: 'The red hot lava was quietly tumbling into the sea over a low ledge, perhaps six or eight feet high, and five hundred to six hundred feet long. The lava did not seem to be quite so liquid, or of such a bright color. . . . It ran more like porridge in great flattened spheroids, which were sometimes partially united together, and sometimes almost separate. . . . There was no steam to be seen escaping from the lava, and it was not until after each spheroidal mass had disappeared for a second or two under water that puffs of steam came to the surface. . . .'

"Mr. Green remarked that this tendency to form spheroids in the molten state might have some connection with the origin of basaltic columns, as well as to weathered spheroidal masses seen in ancient lava streams. . . . He allows that there was nothing like compression: the great flattened spheroids rolled quietly over into the sea, causing a slight commotion in the water."

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GEOLOGICAL NOTES

NATURALLY BRIQUETTED COAL FROM A BED IN WESTERN PENNSYLVANIA[1]

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United States Geological Survey

In the spring of 1945, while I was associated with the United States Bureau of Mines, I received from Dr. Elmer Kaiser, of Bituminous Coal Research, Inc., a very peculiar specimen of coal that was spheroidal in form and about the size of a golf ball. A section across this spheroid showed that granular and pulverized coal had been consolidated around a single larger fragment of coal that served as a nucleus.

The illustrations show the cut surfaces of halves of this nodule, prepared in slightly different ways and photographed, using slightly different modes of illumination. The half shown in plate 1, *A*, has been polished dry, and the surface presents a minimum of relief. Some concentric arrangement of granules is evident, and the nucleus shows normal banding—a feature which is not very evident, however, from this photograph. Two flecks of disseminated pyrite that indicate the plane of banding are more distinctly visible. The specimen was photographed with illumination from a relatively low angle, under which condition polished coal surfaces appear black.

The half that is illustrated in plate 1, *B*, has been gently polished wet on selvyt cloth with optical-grade cerium oxide and shows a central cavity where the nuclear coal broke away. Care in polishing did not prevent development of very high surface relief, and it is clear that the nodule would disintegrate under continued wet conditions. Illumination for this photograph was from a relatively high angle, and the nodule was tilted slightly so that polished coal fragments in the peripheral zone reflected light directly into the camera lens and thus appear white in this picture.

Observed under similar conditions with the Greenough type microscope, the larger fragments of coal appear normally banded, perhaps to a considerable extent microbanded, with dominantly bright luster. A fragment of the peripheral shell was prepared in a thin section at the Bureau of Mines laboratory by Mr. H. J. O'Donnell. The very fragile character of the moistened material occasioned some losses, but an adequate section was obtained for observation. When observed at higher magnification under the biological type microscope by transmitted light, the coal appears mostly vitrinized, much of it of a nature comparable to anthraxylon.[2] It is, however, pulverized to a remarkable degree. Aside from this, it is comparable to bright, high-volatile bituminous coal as in the Lower Kittanning and most of the other coal beds of western Pennsylvania.

Dr. Kaiser transmitted one specimen to me originally, and three additional specimens were subsequently obtained. One of these has been kept intact; the other two were sliced, and they also show an angular coal fragment in the center and a "shell" of pulverized coal with slight concentric layering similar to the specimen illustrated. All have the same type of evenly rounded, slightly roughened dull surface on the outside, with no external cracks or fissures. The only indication of cracks in the shell is seen on the cut surfaces opposite and radiating from the angular corners of the nuclear piece of coal in the center (pl. 1, *A*, and at the back of the central depression of pl. 1, *B*). The nodules are sufficiently coherent that the shell has a tendency to yield shallow, more or less concentric spalls rather than to break across or around the nucleus.

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[1] Text of an informal communication presented before the Geological Society of Washington, March 24, 1948. Published by permission of the Director, U.S. Geological Survey. Manuscript received May 3, 1948.

A portion of the pulverized coal shell from one specimen was submitted to Mr. H. M. Cooper, of the Bureau of Mines Coal Analysis Section, for ash and specific-gravity determinations. The following values were obtained: 17.87 per cent ash, as received; 1.474 apparent specific gravity. The ash appears to be chiefly in the form of finely dispersed clay distributed among the pulverized fragments.

The indirect knowledge of the occurrence in the first place has, unfortunately, prevented a close firsthand appraisal of all the circumstances that might have bearing on the origin of these unusual aggregations. It seems that Dr. Kaiser obtained the specimen he sent me from Mr. Raymond, of the New York Central Railroad, who had received it from inspectors of the railroad, Mr. McHugh and Mr. Cavanaugh, who had, in turn, observed them at the tipple and made some inquiry about them from the mine superintendent, Mr. George H. Henderson. Mr. Henderson had previously noticed the nodules incidentally when they came to the mine tipple, had inquired about their point of origin, and later saw a considerable quantity of them at the pit but had not taken a particular interest in them at that time. When my inquiry finally reached him, this particular pit had been exhausted of accessible coal and was closed. A watch for other occurrences of similar nodules has been maintained since then, but no others have been found. The three nodules sent following my inquiry were discovered by Mr. Henderson at the tipple. Presumably all were from the one original occurrence.

The following details regarding the occurrence are based on a recent interview with Mr. Henderson and an inspection of the locality. The coal is now covered by a considerable amount of talus and could not be directly observed in this old strip pit.

The coal bed in question is the Lower Kittanning. It is about 24 inches thick at this point and overlain normally by some 30 feet of dark-gray shale. The shale is fairly massive-appearing and tough when fresh, though lenticular ironstone concretions are present, but it soon disintegrates on weathering and bleaches to a lighter-gray or buff color. One bony parting is present in the coal, varying from about ½ to 1 inch in thickness and located about 8–10 inches from the top of the bed. This parting was seen at another mine, where it was said to be similar. The bony parting has dull luster, is tough, and varies in general megascopic appearance from a moderately high-ash durain to the tough, shaly, carbonaceous rock characteristically known as "bone." The nodules in question are said to occur *in the coal* for 3 or 4 inches below this bony parting. At the one location where they occurred, the nodules are reported as sticking out of one side of horizontal fracture surfaces of the coal from this layer, leaving a counterpart hemispherical depression on the opposite surface. There is no doubt that the majority of the coal in the bed is normally banded, but a question must remain as to the precise character of the coal inclosing the nodules because no specimen was saved, and at the point where the nodules were found the coal might be different from that normally in this posi-

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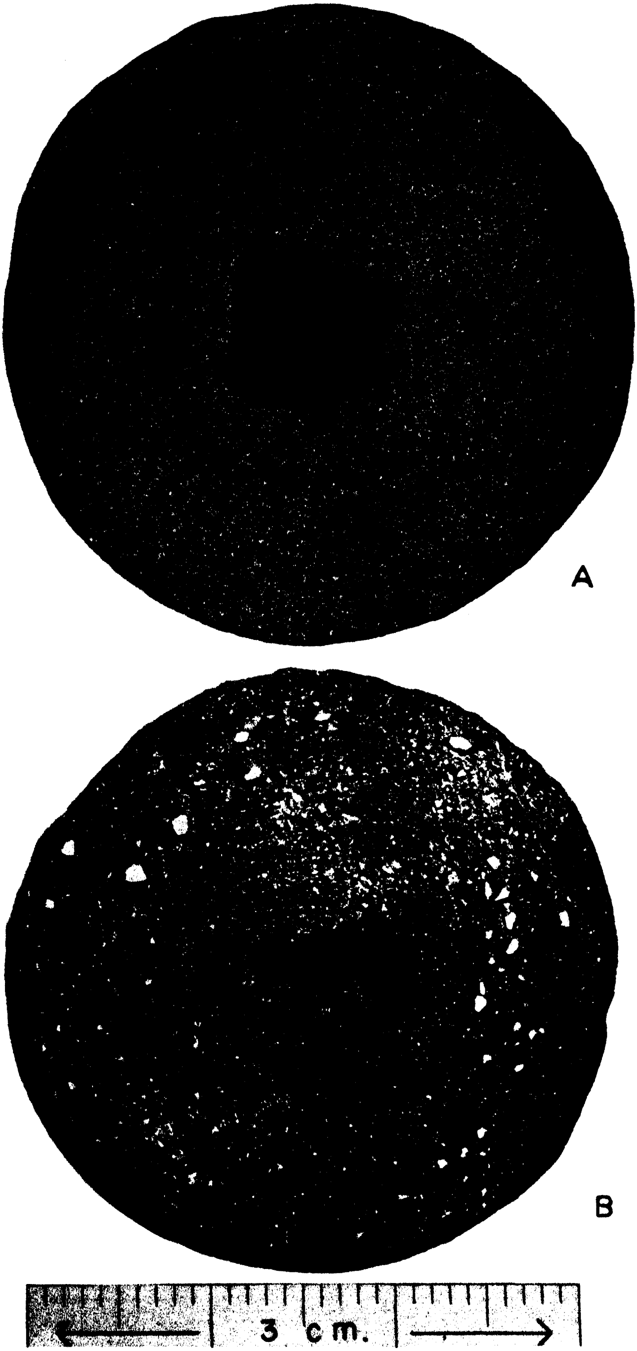
PLATE 1

A, Median section of spheroidal natural briquette, polished dry; angular central coal fragment shows obscure banding (horizontally oriented on the page). Coal fragments appear black.

B, The opposite half of briquette shown in *A*, polished wet, showing considerable relief owing to loss of fine matrix and the nuclear fragment. Larger upstanding polished surfaces of coal fragments appear white in this photograph. Photographs by the author.



Median sections of naturally briquetted coal

tion elsewhere. A normal underclay is said to occur below the coal bed.

The strip pit where the nodules were found is on a small eastward-facing nose of the hill crest just south of Piney Dam on Clarion River. It is located about 3 miles south and 2 miles west of Clarion, Clarion County, in the northwestern part of Pennsylvania. The pit encircles the nose at the level of the Lower Kittanning coal (elevation about 1,400 feet as judged from the 1907 edition of the Clarion quadrangle topographic sheet, reprinted in 1943) to the extent of two or three normal-width "cuts" taken with the electric shovel. The operation was closed when overburden increased to more than 40–50 feet, the present economic limit for the thickness of coal available. The nodules were observed close to the easternmost extension of this strip pit.

The formation of such spheroidal aggregations of pulverized coal is still most puzzling to the writer. The coal composing them was evidently physically similar to bituminous rank coal when they were formed, as shown by the angular fracture edges of the nuclear piece as well as by the angular, comminuted pieces composing the shell. The spheroids are remarkably symmetrical and show no indication of any compression. Not even the internal structure appears to yield any clue in this direction. They are about as hard as lumps of bright coal of the same size would be and less fragile because of the lack of cleating. The ash content of the nodules is considerably higher than average for this bed, and it seems likely that their coherence is partially a result of clay functioning as a "binder" for the pulverized coal fragments. Evidence for this is the tendency of the aggregation to disintegrate under wet grinding and polishing.

These features have implications as to mode of origin, but the suggestions are not easy to reconcile with the well-established usual concepts of coal formation.

It would seem necessary to assume that the coal fragments in the nodules are probably foreign to the Lower Kittanning bed since they must have been physically coalified when the coal they were embedded in was only in the peat stage. It is evident the Lower Kittanning material was easily deformed when the nodules were introduced to this coal bed.

It is obvious that the nodules are not truly concretionary—that chemical processes of solution and reprecipitation, such as are characteristic of mineral concretions, have nothing to do with them.

The aggregation of coalified fragments into symmetrical spheroids could occur only under special circumstances involving rotation in *all* planes around the central nucleus; rotation in even approximate alignment along any single axis would have produced initial axial elongation, a feature that is clearly lacking. One may obtain the type of rotation inferred for the coal spheroids in very special instances of accretionary growth in mud balls or in the organic aggregations Kindle (1934) refers to as "lake balls."[3]

A question must remain as to how the finely pulverized coal of the "shell" could ever have formed under natural conditions. Evidence of the relatively early physical incoalation of beds in Carboniferous times is found in the fairly common occurrence of water-rounded coal pebbles incorporated in the conglomeratic basal zone of channel sandstones in the Pennsylvanian. Such pebbles, like the larger fragments in the nodules, show definite indications of cleating. But the particles and fragments of the accretionary shell of the spheroids described here have no appearance of forming by any processes normal to erosion; they all are sharp and angular and show no evidence of ordinary size segregation or sorting beyond the fact that pieces larger than about 1 mm. were not incorporated in the shell. Smaller pieces of all visible sizes are included, so that whatever size selection existed, apparently acted in reverse of that normally considered affecting size classification. This is, of course, consistent with an accretionary mode of aggregation. Fault gouge material could include similarly unsized angular fragments, but a sufficient supply of coal crushed in this manner and exposed so that numerous accretionary spheroids could be formed at the time of Lower Kittanning coal deposition is difficult to imagine. The writer knows of no collateral evidence that would suggest this remote possibility.

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[3] The coal balls of paleobotanical studies are never accretionary as Kindle suggests; their mineral matrix is truly concretionary.

The lack of secondary compressional distortion of the spheroids by compaction of the Lower Kittanning coal matrix is also hard to understand, yet it must be accepted if the place of discovery is accurately reported, and the writer sees no basis for questioning it. It is possible that some types of attrital coal may have a relatively rapid dehydration-compaction cycle in diagenesis, which is not well understood. The matrix surrounding the spheroids was not examined, and it is possible that it, too, consisted of a pocket of consolidated pulverized coal whose character was not particularly evident on casual inspection. If the immediate matrix consisted of "prelithified" coal fragments, the lack of secondary compression effects in the spheroids would be easier to understand. Under the moderate pressures implied by the present rank of the Lower Kittanning coal in this area, a microbreccia of previously consolidated coal might show imperceptible diagenetic compression like a lens of sandstone.

Enough has been said to give some idea of the highly anomalous character of these spheroids. As the title of this note suggests, the writer regards them as examples of naturally "briquetted" coal because of their obvious but superficial analogy with ordinary commercial briquettes. No similar occurrence is known to be reported.

It is obvious that the natural briquettes have no similarity with the "niggerhead" nodules or coal "pebbles" or "apples," chiefly known now in America from the Walsenburg district in Colorado, though also reported by others (Stutzer, 1940) or observed by the writer from the anthracite and southwestern districts of Pennsylvania, near Red Lodge in Montana, in Alaska, Peru, Poland, Hungary, France, northern Italy, India, the Transvaal, and Australia. These arise from an unusual type of spheroidal spalling consequent to normal rectangular cleating, and close observation shows that there is no derangement of the coal in the bed. Coal balls are essentially a type of true mineral concretion in coal beds and, likewise, have no bearing on or relation to the spheroidal "briquettes" described here. Sometimes large examples of vertical conchoidal cleating (eye-coal, "augen"; Stutzer, 1940, pp. 249–253) in coal of homogeneous lithology have been mistaken for spheroids of one kind or another broken in half, but these features must be related to a very different type of phenomenon. The manner of origin of the natural briquettes is another of the many unsolved mysteries about coal.

The specimens described in the foregoing communication have been deposited for reference in the mineral collections of the United States National Museum under catalog no. 108467.

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STYLOLITES IN PRE-CAMBRIAN QUARTZITE[1]

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INTRODUCTION

The stylolites here discussed came to the writer's attention during the petrographic study of a suite of gneissic rocks from the Ace Lake area, some 8 miles north of Goldfields, Saskatchewan. Alcock (1936) included these rocks with granite and gneisses in one mappable unit, which he considered to be of Archean age, younger than the Tazin group and older than the Beaverlodge series of early Proterozoic age. Recent work by A. M. Christie,[2] of the Geological Survey of Canada, has cast some doubt on the validity of separating the Tazin group and the Beaverlodge series. The metamorphic rocks in the Ace Lake area may therefore be referable to either the Tazin group or the Beaverlodge series.

ORIGIN OF STYLOLITES

The question of whether stylolites are of primary or secondary origin has been widely discussed. There is a difference of opinion even with reference to the same occurrence of stylolite seams in the Pottsville quartzite of Lookout Mountain, Alabama. Shaub (1947, p. 116) believes them to be "entirely due to the usual processes of deposition of cross-bedded sands together with the adjustments occurring during compaction up to the time of the consolidation of the sandstone." On the other hand, Stockdale (1945, p. 135) states:

> Instead of being developed as primary structures along normal planes of sedimentation at or near the time of deposition of the original sediment (sand), the stylolite-seams came into being as secondary features subsequent to the original deposition and compaction of the sediments. They were developed along fracture-partings which originated at a time not only later than the original consolidation of the sediments into sandstone but even later than the metamorphism into the quartzite. Thus there is revealed another finding to be added to the already existing preponderance of evidence in support of the belief that stylolites are of secondary origin as held by the solution theory.

According to Stockdale (1943, p. 3):

> The *solution theory* contends that stylolites result from differential chemical solution in hardened rock, under some pressure, on the two sides of a parting of some sort (such as a bedding plane, lamination plane, suture, or crevice), the individual portions of the one side fitting into the dissolved out parts of the opposite, the interfitting taking place slowly and gradually as solution continues.

In referring to stylolites in the Pottsville sandstone of Pennsylvanian age and the Medina sandstone of Silurian age, Price (1934, p. 188) expressed the view that they were due to solution of material along parting surfaces and that pressure was not a factor; for he states: "In the present examples differential solution alone can account for their origin." In contrast to this view, the hypothesis of a primary origin is advocated by Shaub (1939, p. 47), who states: "The occurrence of stylolites in sandstone and quartzite, as well as in the carbonate rocks, is readily accounted for by the contraction-pressure theory, which also explains the wide variation in form, structure, markings and other features of stylolites."

The purport of this paper is not to suggest that *all* stylolites are of secondary origin but to demonstrate that in this instance the origin could not have been primary.

DESCRIPTION OF THE PRE-CAMBRIAN QUARTZITE

The rock is a fine-grained, dark-gray quartzite, with a network of minute, quartz-filled fractures. As can be seen in plate 1,

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[1] Manuscript received May 3, 1948.

[2] Personal communication.

the stylolite is more or less parallel with the foliation, which, in transmitted light, is indicated by wavy stringers of specularite and, under crossed nicols, by the elongation of the quartz grains. The grains of quartz vary in length from 0.03 to 0.1 mm. and have sutured boundaries. The elongation of the quartz grains and the sutured boundaries indicate that the rock has been recrystallized under stress. The stringers and surfaces of the stylolites are of minute specks and clusters of specks of specularite, bluish-gray by reflected light.

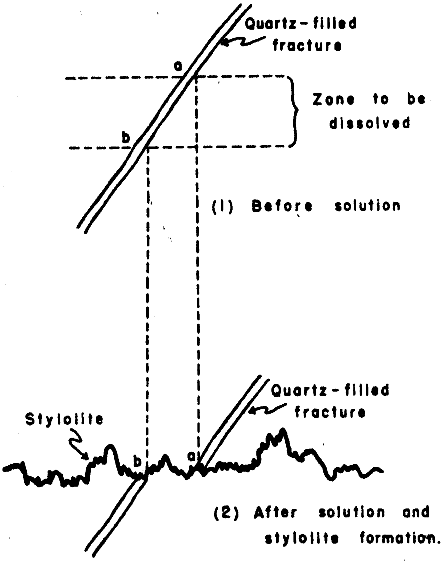

FIG. 1.—Apparent displacement of a quartz-filled fracture by a stylolite, due to solution along the cross-cutting stylolite.

RELATIONSHIP OF THE STYLOLITES TO THE QUARTZITE

An examination of the thin section illustrated by plate 1 shows that the stylolites must have developed after the recrystallization of the rock. The following points indicate the age relationship of the stylolites to the quartzite and specularite stringers.

1. The quartzite has been formed by recrystallization under stress, as indicated by the sutured quartz grains elongated parallel to the foliation. If the stylolites were of primary origin, they would have been destroyed.

2. All the quartz-filled fractures cut the specularite stringers. Furthermore, the fracture illustrated in plate 1 is a minute fault, as indicated by the displacement of the band of coarser quartz beneath the stylolite.

3. The stylolite shown in plate 1 cuts the quartz-filled fracture and must have originated after the period of fracturing.

4. The stylolites were not all formed at the same time, as several of the quartz-filled fractures cut stylolites, although they are themselves cut by other stylolites.

5. Where stylolites obliquely cut quartz-filled fractures, there is apparent displacement of the fractures. This apparent displacement is not considered due to movement but to the effect of solution along the cross-cutting stylolites (fig. 1).

An examination of plate 1 shows that the stylolite forms one boundary of an area of clear quartz and that other irregular lenses of clear quartz are parallel to the foliation. Plate 1, *B*, shows that the quartz grains in these clear areas are coarser than are those associated with the specularite stringers. In the adjacent rocks specularite has been introduced in abundance, and it is inferred that the specularite stringers have also been introduced and do not represent the iron content of the original sediment. Unless the finer-grained quartz offered better channels for the infiltration of iron-bearing solutions,

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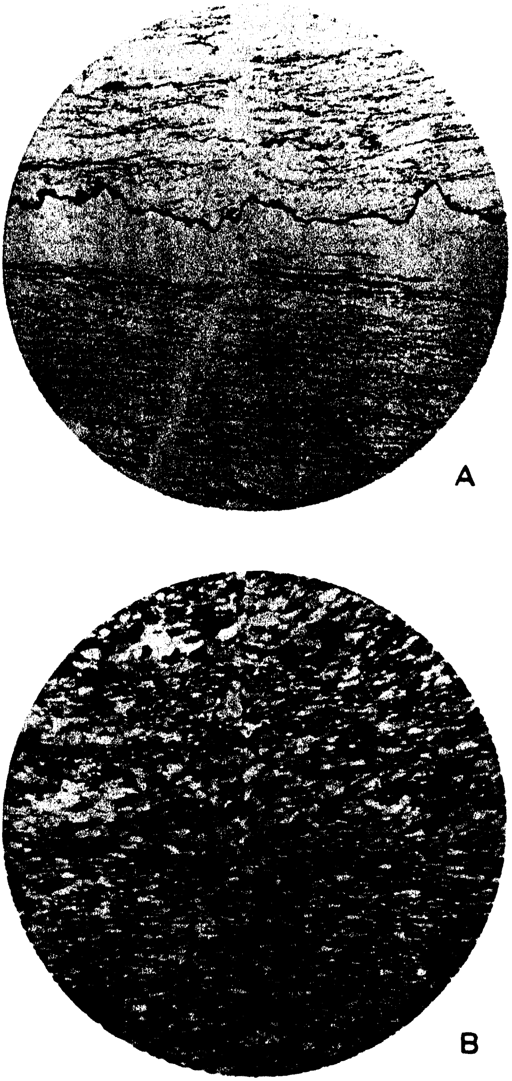
PLATE 1

A, Thin section of quartzite, viewed with transmitted light. The dark stringers are of specularite. Note that the quartz-filled fracture cuts the specularite stringers and is itself cut by the stylolite.

B, Same view as *A*, between crossed nicols. Note that the irregular lenses of clear quartz and the area of clear quartz adjacent to the stylolite are coarser grained.



Stylolites in pre-Cambrian quartzite

it must be inferred that the areas of clear quartz are of later development than the specularite. If so, the quartz may have been introduced or may have been locally recrystallized. Petrographic examination of adjacent rocks shows that similar quartz lenses and stringers have been introduced, and the same sequence is inferred here. As is shown by plate 1, the quartz-filled fracture cuts these areas of clear quartz.

It is inferred that the stylolites have developed by solution of the rock along partings, accompanied by the concentration of residual specularite along the surfaces of the stylolites.

CONCLUSIONS

The stylolites here described are of secondary origin and were formed after the recrystallization of the quartzite. It is demonstrated that the sequence of events in the rock's history has been as follows: First, there has been recrystallization of the original rock to a quartzite, the recrystallization having taken place under conditions of stress, as indicated by the parallel orientation of the sutured quartz grains. Second, fracturing of the rock has occurred transverse to its foliation, quartz having been introduced into the fractures. Third, there has taken place the development of those stylolites which cut the quartz-filled fractures.

In conclusion, the writer wishes to state that, while the mode of origin of the stylolites here discussed is in doubt and can only be inferred, the time of origin is unquestionably later than the period of recrystallization of the rock.

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RECENT SHORELINE CHANGES AT SHIRLEY GUT, BOSTON HARBOR[1]

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INTRODUCTION

Prior to 1936, Deer Island in Boston Harbor was separated from the mainland by Shirley Gut, a strip of water with unusually strong hydraulic currents (fig. 1). This gut was filled in by the growth of spits from Deer Island and Point Shirley.

Point Shirley, a part of the mainland, consisted of a small drumlin, from which a broad spit extended southeastward (LaForge, 1932, pl. 2). Deer Island was composed of four eroded drumlins joined by beaches. Until 1936, a spit at its northwest end terminated at Shirley Gut (Roorback, 1910).

COLONIAL AND RECENT HISTORY OF SHIRLEY GUT

There are records of boats sailing through Shirley Gut in early Colonial times. It was used in the Revolutionary War as the northern entrance to the harbor, and in 1776 a fort was built on Point Shirley to defend it. On May 19, 1776, a naval engagement took place in the Gut between the colonists and

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[1] Manuscript received March 4, 1948. Published by permission of the Director, Geological Survey, U.S. Department of the Interior, and the Massachusetts Commission of Public Works.

the British. The colonists had two boats—the privateer "Franklin," of four guns and about thirty men, and the "Lady Washington," a schooner carrying twenty-five men. A strong force of more than two hundred British in small boats attacked the schooners. The colonists overturned many of the British boats, and the men who were plunged into the waters of the Gut were carried away from the scene of the battle by the swift currents. The British were repulsed and are said to have lost seventy men and several boats, but the colonists lost only one man. Up to about 1900 the Gut was used by boats of two or three hundred tons, some of which carried passengers to Nahant and elsewhere. Since 1900 the Municipal House of Correction has been on Deer Island, and the Gut has served as an effective barrier against escape, for the currents are so powerful that only the strongest swimmers could reach the mainland.

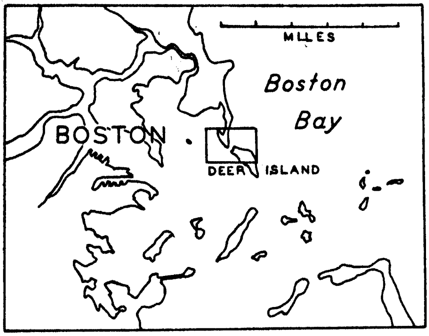


FIG. 1.—Index map of Boston Harbor, Massachusetts. Area considered in text outlined.

CLOSING OF SHIRLEY GUT

The growth of the Point Shirley and Deer Island spits, as well as the filling-in and destruction of Shirley Gut, is shown in figures 2, 3, and 4. In 1860 the average width of Shirley Gut was 700 feet. By 1898 its width had decreased to 400 feet; in 1934 it was only 80 feet wide at its narrowest point; and in 1936 Deer Island was joined to Point Shirley by a strip of land which was 300 feet wide at mean tide and completely flooded only by exceptional tides (figs. 2 and 3). In places the Point Shirley spit grew approximately 400 feet in the period between 1860 and 1934. Before 1936 the northwestern end of Deer Island was a double-pronged spit. The bayward prong on the south side of the spit was relatively inactive between 1860 and 1934, although in places the oceanward prong grew nearly 750 feet during the same period.

While these spits were growing toward each other, the Gut was also being filled in (fig. 4). In 1847 the maximum depth of the Gut at high tide along the profile shown in figure 4 was 45 feet. This depth was due partly to scour. In 1861 it was 35 feet, and in 1934 about 11 feet. An analysis of the profiles shows that the filling did not take place uniformly, as there were both cut and fill.

CONFLICT BETWEEN HYDRAULIC AND LONGSHORE CURRENTS

During the existence of the Gut there was a continual conflict between the hydraulic currents running through it and the longshore currents which formed the spits. The former tended to keep the Gut open, whereas the latter, in general, tended to fill it. The strength of the hydraulic currents depended mainly upon the tidal range and the force and direction of the wind, and the strength of the longshore currents depended upon the force and direction of the waves. The strengths of both the hydraulic and the longshore currents varied greatly with continual changes in the direction and force of the wind and waves and with changes in the tidal range. It is not surprising, therefore, that the filling of the Gut did not proceed at an even pace. As the Gut narrowed, the currents in it became stronger, so that the tendency to fill diminished. Just before the joining of Deer Island and the mainland, however, the volume of water going through the Gut was decreased; the erosive effect of the currents was consequently diminished, and the filling proceeded rapidly.

The currents at the bottom of Shirley Gut had velocities at times in excess of 1.7

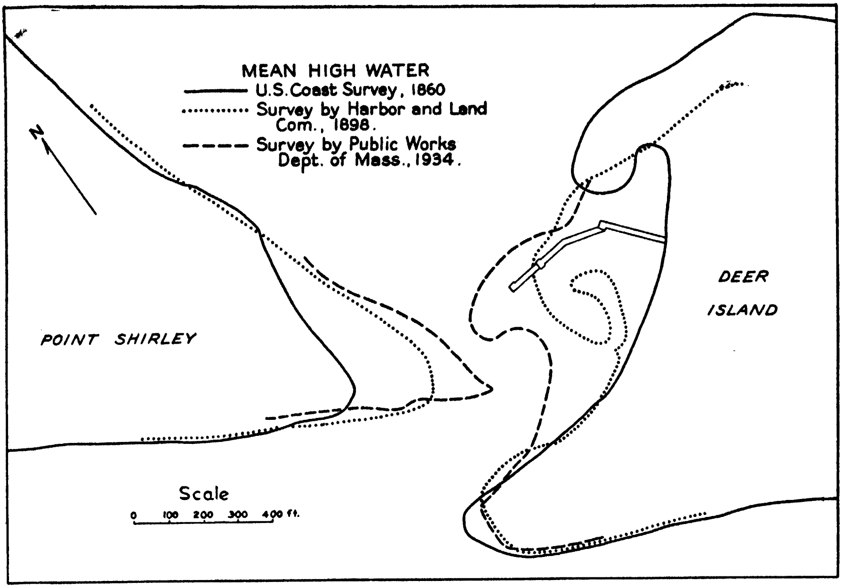
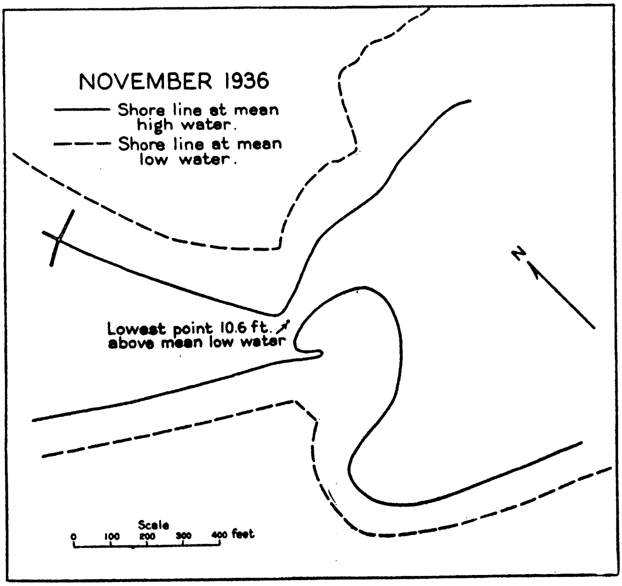


FIG. 2.—Map of Shirley Gut in 1860, 1898, and 1934


NOVEMBER 1936
Shore line at mean high water.
Shore line at mean low water.
Lowest point 10.6 ft. above mean low water
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Scale
0 100 200 300 400 feet

FIG. 3.—Map of Shirley Gut in 1936. Map by R. L. Nichols and Louis Riseman

knots,[2] equal to 0.87 meter per second. These velocities could have moved fragments larger than a bean or hazelnut (Twenhofel, 1932, pp. 40–41). The material which was carried to the Gut by the longshore currents ranged in size from silt to boulders. The small sizes were removed by the hydraulic currents in the Gut, but the boulders and cobbles, being too heavy to be moved, accumulated in the Gut. Soundings made before the Gut closed showed that it had a gravel bottom.

GROWTH OF DEER ISLAND SPIT

The growth of both the Deer Island and the Point Shirley spits was responsible for the joining of Deer Island to the mainland. The north prong of the Deer Island spit prograded more rapidly than did the Point Shirley spit prior to the closing of the Gut. This is indicated by the following facts: (1) The data afforded by figure 2. (2) Before the closing of the Gut, the material of Point Shirley Beach systematically varied in size from place to place.[3] At Winthrop Head, which is north of Point Shirley (fig. 1), the beach was composed of coarse gravel; southward the beach material gradually decreased in size; and, still farther south and extending to the end of Point Shirley, it gradually increased in size. The coarse material at the end of Point Shirley was drifted across the Gut from Deer Island. This could have been accomplished only if the currents from Deer Island to Point Shirley were, in the main, stronger than those moving in the opposite direction. This fact suggests that the growth of the Deer Island spit was more important in closing the Gut than was that of the Point Shirley spit. (3) Immediately prior to the closure, the Deer Island spit was advancing over the Point Shirley spit (Nichols, 1938).

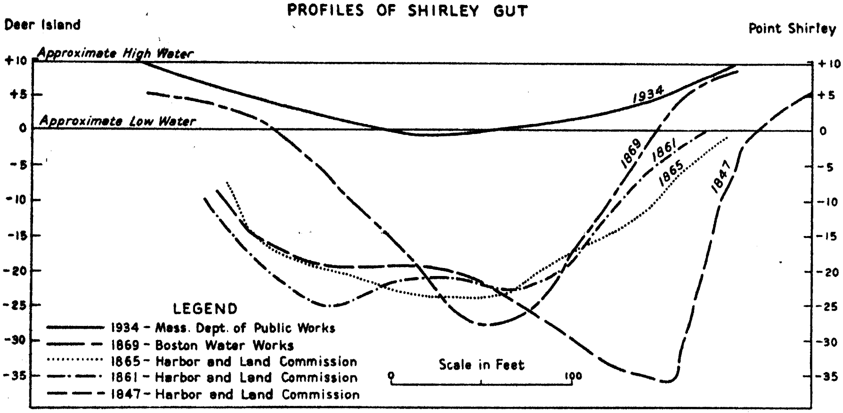

FIG. 4.—Transverse profiles of Shirley Gut

The closure of Shirley Gut, as is indicated by figure 2 and an analysis of the longshore currents, resulted from the movement of material along the ocean side of the two spits. After the closure was completed, a small spit grew southeastward from the bay side of the former Point Shirley spit (center, fig. 3). If man had not interfered with the normal shoreline development, this small spit in time would have reached the former Deer Island spit, and a small pond would have been formed.

ACKNOWLEDGMENTS.—Thanks are due to Dr. L. W. Currier, of the U.S. Geological Survey, for help in the preparation of the manuscript, and to Messrs Theodore Botinelly and Louis Riseman, both of Tufts College, for assistance in the mapping.

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[2] L. P. Disney, personal communication, 1942.

[3] A. C. Lane, personal communication, 1936.

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ROCKS OF THE MID-ATLANTIC RIDGE[1]

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A remarkable achievement by the Atlantis expedition (1947) of the Woods Hole Oceanographic Institution, under the leadership of Professor Maurice Ewing, was the recovery of several hundred pounds of boulders from depths approaching 2,500 fathoms over the Mid-Atlantic Ridge. Some of these boulders are of limestone and will be described elsewhere, but a large proportion consists of eruptive rocks referable to the following groups: (1) holocrystalline gabbroic rocks, (2) basalts (pillow lavas) with and without olivine, and (3) serpentines. Table 1 gives the latitude and longitude of each station where these boulders were collected, together with the depth and the kind of rock.

PETROGRAPHY

GROUP 1—HOLOCRYSTALLINE GABBROIC ROCKS

Station 5, no. 1.—This is a fresh olivine gabbro of grain size $\frac{1}{2}$–1 cm. It contains hypidiomorphic to rounded grains of olivine, sometimes with narrow rims of talc (?) and magnetite. The olivines are in many cases enclosed in larger individuals of clinopyroxene (augite with polysynthetic twinning). The plagioclase is a medium labradorite having an imperfect ophitic relation toward augite.

Station 6, no. 3.—This rock may be described as a crushed anorthositic gabbro. Corroded plates and prisms of diallagic pyroxene, some as much as 2 cm. long, are surrounded by collars of granulated material of the same appearance. These aggregates lie in a fine, sugar-grained groundmass of plagioclase, within which the pyroxene forms dark streaks and eyes. The larger pyroxene fragments have a micaceous cleavage, and the cleavage cracks are strongly curved. In consequence of this curvature, with the accompaniment of wavy extinction and minute polysynthetic twinning, it was found impossible to make any reliable measurements of optical orientation or axial angle. The same difficulty recurred in the case of the plagioclase grains; in many of these the polysynthetic twinning is faint or almost absent, and in the few grains that seem to show distinct lamellae, it appears, when the grain is examined on the universal stage, that each lamella is made up of a great number of thinner lamellae of which no measurements could be made. All that one can say of this plagioclase is that the mean refractive index is well above that of Canada balsam. The grains of the granulated pyroxene are mostly between $\frac{1}{10}$ and $\frac{1}{100}$ mm. They fit closely together, and films of brown biotite lie along the junctions. The larger plagioclase grains are of the same dimensions, but where crushing has been most intense the smaller grains have been so closely compacted that the boundaries are not sharply discernible.

Station 20, no. 6.—This specimen is a coarse and remarkably tough diabase, in which plagioclase laths 1–3 mm. long are enclosed in skeleton crystals of augite several times that size. Little, rounded grains of olivine, transforming externally into iddingsite, are also present. This rock is a perfectly normal olivine diabase.

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[1] Manuscript received July 19, 1948.

GROUP 2—BASALTS WITH AND WITHOUT OLIVINE

Station 7, no. 1.—The specimen is a fine-grained gray rock with a few little insets of plagioclase and olivine. One end of the specimen shows a distinct curvature and bears a skin of shiny black glass. Perpendicular to this skin, the rock shows vestiges of a radiating columnar structure, indicating that the specimen is part of a lava pillow. A thin section was cut about 6 inches beneath the glassy skin. Under the microscope it showed about 5 per cent of olivine insets, measuring 0.25 mm. or less and remarkable for their perfect freshness. A network of thin plagioclase laths ranging in length from 1.5 mm. to the limit of visibility, lies in an imperfectly crystallized groundmass consisting of augite granules, feldspar microlites, and semiopaque matter. There is no glass. The extinction of the plagioclase laths indicates a rather calcic labradorite.

Station 7, no. 3.—A thin section, cut about 3 inches beneath the glassy crust of this specimen, is generally similar to no. 1 but less completely crystallized. There are many little idiomorphic insets of olivine, holding oriented inclusions of the groundmass. The latter consists of radial groups, brushes, and sheaves of very thin plagioclase laths, separated by, and to some extent including, some semiopaque dark glass or near-glass.

TABLE 1

Station	Latitude	Longitude	Depth (Fathoms)	Rocks
5	30°00′	42°10′	2,100–2,500	Single specimen of olivine gabbro
6	30°06′	42°08′	800	Crushed anorthositic gabbro; much serpentine; no basalt
7	30°01′	45°01′	2,340	Basalt only
8	31°49′	42°25′	1,700–2,100	One fragment of weathered basalt
10	31°55′	40°26′	1,450–1,500	One sample of basalt
20	34°04′	42°16′	2,250	Much basalt; much serpentine; one coarse diabase
21	30°08′	43°37′	2,300	Much serpentine; no basalt

Station 7, no. 5.—A thin section of the glassy crust of this sample shows a perfectly isotropic, clear yellow-brown glass containing little insets of olivine and plagioclase, but no augite. Each of the plagioclase laths is surrounded by a mantle of darker-colored glass, which, under high magnification, seems to have a fibrous structure. It was not possible, however, to determine the nature of these mantles with certainty. There are also many minute or incipient crystals in the glass, each surrounded by a dark mantle. The olivine crystals show no such mantle. Fuller (1932, p. 107) has described similar dark mantles about crystals of plagioclase and augite in a basaltic glass (sideromelane) from the Columbia River basalts. He considers the mantles to represent "a transitional step in the development of an opaque ground composed of dustlike particles and thin needles of magnetite." It is self-evident that the crystallization of feldspar from an iron-containing glass must enrich the surrounding glass with iron, while the crystallization of olivine uses up iron and therefore generates no dark mantle.

Station 7, no. 7.—This rock is identical with no. 1.

Station 7, nos. 4 and 6.—These specimens are so similar to those described above that they were not studied under the microscope.

Station 8, no. 1.—This fragment is brown and much altered. It is again an olivine basalt, but the olivine is largely altered to an iddingsitic material. The groundmass contains spongy iron hydroxide, and vesicles are lined with bright-green chlorite or filled with fibrous zeolites.

Station 20, no. 2.—In this specimen, insets of glassy plagioclase, up to ½ cm. in diameter, and a few small augites lie in a dark-gray, fine-grained base with many minute vesicles. On the universal stage the augites show an axial angle of 52–54 (positive). The indicated composition of two albite-carlsbad twins of plagioclase was Ab_{65} and Ab_{70}, respectively. Olivine is also present among the insets. The groundmass consists of little plagioclase laths and some olivine grains, the interstices filled with a semitransparent, cryptocrystalline matrix. All the insets are perfectly fresh.

Station 20, no. 3.—This basalt seems to be free from olivine, not even represented by serpentine or iddingsite. Abundant thin laths of plagioclase, containing inclusions of the groundmass, form radiating groups and brushes in a rather indistinct ground of smaller plagioclases and little prisms and grains of augite, with small amounts of calcite and chlorite. There is no true glass.

Station 20, no. 4.—A very fine-grained, dark-gray rock without visible insets of any kind. Under the microscope a few platelets of plagioclase may be seen, but the bulk of the rock is spherulitic. Neither olivine nor augite has crystallized.

Station 20, no. 7.—A rather unfresh basalt with semiophitic texture. No olivine seems to be present, although the matrix contains greenish decomposition products, in addition to augite and plagioclase.

GROUP 3—SERPENTINES

Station 6, no. 1.—Antigoritic serpentine has almost completely replaced original olivine, of which only a few scraps survive. Large plates of bastite presumably represent an original pyroxene. The rock is cut by veinlets of chrysotile. Iron oxides, talc, and carbonates are other secondary products.

Station 6, no. 2.—This specimen is like no. 1, but the olivine is completely replaced.

Station 20, nos. 1 and 5.—In addition to varieties of serpentine, these rocks contain tremolite and a few grains of brown spinel (chromite?).

Station 21, no. 1.—This specimen is a mass of tremolite asbestos.

Station 21, no. 2.—This rock is another bastite serpentine with some residual olivine.

DISCUSSION

Our knowledge of the composition of the Mid-Atlantic Ridge has been derived hitherto from the islands and island groups which are scattered at long intervals along its course, namely, Gough Island, Tristan d'Acunha, St. Helena, Ascension, St. Paul's Rocks, and the Azores. On every one of these islands the dominant rock is basalt, generally olivine-bearing but sometimes free of this mineral. Other rocks which have been recorded are as follows:

Gough Island.—Essexite, trachyte, sodalite-trachyte, phonolite (Campbell, 1914; Smith, 1930; Barth, 1942).

Tristan d'Acunha.—Pyroxenite, trachyte, phonolite, single boulder of muscovite-biotite gneiss or granite (Schwarz, 1905; Smith, 1930).

St. Helena.—Trachyte, phonolite (Daly, 1927; Smith, 1930).

Ascension.—Trachyte, rhyolite, obsidian, boulders of granite, diabase, gabbro, peridotite (Darwin, 1876; Renard, 1889; Reinisch, 1912; Daly, 1925).

St. Paul's Rocks.—Mylonitized dunite (with hornblendic and enstatitic facies) (Renard, 1879, 1882; Washington, 1930; Tilley, 1947).

Azores.—Trachyte, essexite (John, 1896; Reinisch, 1912; Friedlaender-Esenwein, 1929).

The Atlantis collection confirms the predominance of basalt and makes at least two notable additions to the record, namely, an abundance of serpentine and a mylonitized anorthositic gabbro. The latter, together with the dunite-mylonite of St. Paul's Rocks, provides evidence of strong shearing stresses during or after the formation of the Ridge. The serpentines show advanced hydration and other degenerative changes, whereas the olivine-basalts are just as fresh as random samples of any recent terrestrial basalt and contain unaltered olivine. Thin veinlets of serpentine cut the St. Paul dunite, but the very slight degree of serpentinization in this rock and in the olivine-basalts contrasts strongly with the almost complete hydration exhibited by the serpentines. This seems to justify a suggestion that the serpentine may be appreciably older than the basalts.

Professor R. A. Daly has shown by comparison of chemical analyses that the basalts of Ascension Island differ in no essential respect from the average basalt of the continents. He concludes that the similarity of these averages illustrates "the relative uniformity of common basalt throughout the world" (1925, p. 73). The basalts of the Atlantis collection support that conclusion. They show no mineralogical or textural characters that cannot be duplicated among terrestrial basalts.

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DISCUSSION

RADIAL DIFFUSION AND CHEMICAL STABILITY IN THE GRAVITATIONAL FIELD: A DISCUSSION[1]

JEAN VERHOOGEN
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The interesting paper by Mr. Ramberg which appeared in the September, 1948, issue of this *Journal* contains several statements which may require further elaboration because they might lead geologists who are unaware of certain thermodynamic principles into erroneous conclusions. I have in mind particularly the effect of pressure on the stability of hydrous minerals (p. 455). While discussing the stability of minerals in a gravitational field, Mr. Ramberg should perhaps have pointed out, as is well known (Guggenheim, 1933), that chemical equilibrium is independent of position in such a field. If a system is in a given state of equilibrium in a closed container, this equilibrium remains unchanged whether the container is lifted 100 miles in the air or is lowered to the center of the earth, provided that the pressure in the container remains constant. When rock masses change position in the earth, the pressure acting on them does not usually remain constant but varies according to the hydrostatic law; it is this change in pressure which is responsible for such shifts of equilibrium as may result from the change in position. The effects of these changes in pressure are the same as those of any other change in pressure. In particular, the shifts of equilibrium resulting from an increase in depth must be such as would result in a decrease in the total volume of the system.

Thus, if we consider a reaction of the type

$$A \rightarrow B + H_2O ,$$

where A is a hydrous mineral and B an anhydrous one (or several), an increase in pressure will result in the formation of B at the expense of A only if the sum of the molar volumes of B and water is less than that of A. The increase in pressure, dP, required to start the reaction as written must be such that

$$|\Delta V dP| \geqslant |\Delta G| ,$$

where ΔV is the change in volume resulting from the reaction and ΔG is the free energy of formation of A from B and water at the reference pressure. Because the molar volume of water is sensitive to changes in P and T, no general statement can be made concerning the magnitude of the pressure increase required to start the dehydration reaction; in some cases this increase might be greater than the pressures encountered at the depths where metamorphic rocks are formed. Mr. Ramberg's statement that "high temperature is not necessary for the formation of the anhydrous minerals of the granulite facies" is perhaps correct under certain circumstances; it cannot be so under conditions in which the molar volume of water is large, for in such a case the effect of pressure alone would be toward the formation of the hydrous mineral at the expense of the anhydrous one.

In the same manner the "squeezing-out" of large ions by pressure may occur only if these ions, after they have been squeezed out, go into some form of lesser volume. Mr. Ramberg apparently takes the view that they move out of the system, so that their volume is not to be taken into account at all. This could happen if the squeezed crystal could evaporate into a vacuum or a medium of very low pressure, a case which is probably not encountered commonly in the depths of the earth. The squeezing-out mechanism may work in some cases; it may not work under different conditions. In each case it should be clearly specified what the conditions of pressure and temperature are assumed to be; and also the form under which the squeezed-out ions are supposed to leave the original lattice should be stated, so that the volume relations can be worked out. But even if the volume relations are right, the minimum pressure required to start the squeezing will still depend on the energy with which the ions are held in their original lattice and may in some cases be greater than the pressures encountered under terrestrial conditions.

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[1] Manuscript received October 13, 1948.

The statement concerning wollastonite at the top of page 456 may perhaps also require clarification. The present writer is not sure that he understands what is meant by "at ideal gravitational stable conditions"; but he gathers from the context that Mr. Ramberg has in view the case of a vertical column of density ρ, in which the pressure increases with depth h as $dP_1 = \rho g dh$, in contact with a column of CO_2 in which the pressure increases as $dP_2 = \rho_{CO_2} g dh$. If both columns are at the same uniform temperature and if the pressure of CO_2 at the top of the column is such that equilibrium exists there with regard to the reaction

$$CaCO_3 + SiO_2 \rightleftharpoons CaSiO_3 + CO_2\,,$$

then the equilibrium will be displaced at the bottom of the column toward the formation of wollastonite only if

$$(v_{CaCO_3} + v_{SiO_2})dP_1 > v_{CaSiO_3}\,dP_1 + v_{CO_2}\,dP_2\,,$$

where v stands for molar volume. Substituting numerical values, we find that ρ must be greater than 2.24 gm/cm³, or, in other words, the calcite–quartz-bearing rock must have a density greater than 2.24, which will usually be the case. But in the usual case, also, the increase in pressure on the CO_2 will be the same as that on the minerals; wollastonite would then form at the bottom of the column only if the density of CO_2 exceeded 2.24 gm/cm³. Thus, at least near the surface, the equilibrium would be displaced at depth toward the formation of calcite, as usually stated.

There are a few other points in Mr. Ramberg's paper to which one might take exception, from a thermodynamic point of view, for example, his synonymous use of the words "equilibrium" and "stability." The statement in the second paragraph at the top of page 450 appears ambiguous to the present writer: surely, Mr. Ramberg does not believe, as he appears to do, that the relation between chemical potential and vapor pressure, which is really in the nature of a definition, changes with position in the gravitational field. It is, of course, precisely because the chemical potential changes with position in a gravitational field that equilibrium can be maintained only by compensating pressure and concentration gradients. The matter is clearly presented by Guggenheim (1933, pp. 153-159). The reader will find there, in a more rigorous form, the equations which Mr. Ramberg derives for the changes in composition in a binary ideal solution. Also, we surely did not have to wait for Mr. Ramberg's "proof that mechanical instability in the gravitational field results in chemical or thermodynamic instability"; the whole matter was dealt with adequately by Gibbs seventy years ago.

The above discussion should not detract from the merits of Mr. Ramberg's contribution. The present writer's purpose is mainly to emphasize how difficult it is to word a statement that will remain correct under all circumstances, although it may be correct under certain specified conditions. Because most petrologists are not in a position to follow the intricacies of thermodynamics, it should be pointed out to them that there is usually more than one side to a question. More specifically, the conditions under which a statement is believed to be true should be carefully stated, in order to avoid unwarranted extrapolations.

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RADIAL DIFFUSION AND CHEMICAL STABILITY IN THE GRAVITATIONAL FIELD: A REPLY[1]

HANS RAMBERG
University of Chicago

Everyone must agree with Mr. Verhoogen that a change of position in the gravitational field does not displace chemical equilibrium in a system closed in an impermeable container. It is equally obvious—as Mr. Verhoogen discusses at length—that if such a closed system is moved up and down in the earth, then pressure is the only variable which can displace chemical equilibrium within the system (assuming a constant temperature). The effect of pressure upon chemical equilibria in closed systems is adequately expressed by Clapeyron's equation (so commonly cited in textbooks on petrology).

These facts are, however, irrelevant to a discussion of my paper on chemical stability in the gravitational field, which necessarily implies changes of bulk chemical composition at different depths in the earth. In other words, only rocks of certain compositions are theoretically stable at given depths. In order to determine this theoretical relationship between depth and bulk chemical composition, one must assume that the rocks are more or less open systems.

By introducing the concept of a small closed system, which is moved up and down in the crust, the crucial points in my article cannot be discussed at all because the chemical communication between the system and its surroundings is artificially hindered by a completely impermeable box.

By "open system" I do not mean that the rocks necessarily are porous and full of fissures. The term should indicate only that substances are able to migrate between the system (rock) and its surroundings.

In order that a certain type of particle (atoms, ions, or molecules) may diffuse out of a rock under high pressure, it is not necessary that the rock be surrounded by a "vacuum or other medium of low pressure." The necessary thermodynamic requirement for diffusion in a certain direction is that there exists a gradient in the partial chemical potential in the same direction. The chemical potential, for example, may be high at low mechanical pressure and low at high pressure, and it follows that diffusion in such a case will be from low to high mechanical pressure. For this reason water molecules diffuse from low to high pressure in osmotic systems. Large porphyroblasts can grow in rocks under high confining pressure for the same reason.

It is well known that the thermodynamically stable state of a system is independent of the ways and velocities of reaction and can be calculated without knowledge of the kinetics of the processes. Likewise, the spontaneous reactions taking place in an unstable system tend to achieve the stable state. It may thus be worth while to try to consider the whole earth—or at least larger parts of it—as enormous and complicated physicochemical systems in which there exist instabilities of different kinds. The processes that we observe tend to change the earth as a whole into more stable states, although it is very unlikely that complete thermodynamic stability will ever be reached (C. J. Sullivan [1948] has recently published an interesting paper in which similar views are applied to granitization and ore formation).

The mechanical instabilities may be reduced by mechanical readjustment or bodily motion of matter. The chemical instabilities will usually be reduced by chemical processes only. Adjustment of large-scale chemical instabilities requires, therefore, chemical processes which are identical with long-distance diffusion.

Two questions arise: Do large-scale chemical instabilities exist in the earth? If so, does adjustment of these instabilities take place? A positive answer on these questions implies that large-scale diffusion is of geological importance. A negative answer on the second question simply means that geologists need not concern themselves with large-scale diffusion at all.

In the article discussed by Mr. Verhoogen I did not aim to point out all the large-scale chemical instabilities which exist in the earth or to show what the completely stable state of the earth should be. My aim was merely to show how gravity alone affects chemical stability in the globe. I am well aware of the fact that the effect of gravity is commonly overshadowed by the effects of other factors, such as differences of pressure or temperature or intermingling of chemically incompatible rocks. On the other hand, only by complete knowledge of the process

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[1] Manuscript received October 25, 1948.

of gravitative chemical differentiation shall we be able to assign to this particular phenomenon its proper place in the whole geological picture. Therefore, considerations of the chemical processes taking place within small impermeable boxes moved up and down in the crust do not seem to be very fruitful in the solution of these questions. In one example Mr. Verhoogen touches the problem of large-scale chemical stability in the gravitational field: He notes that wollastonite tends to form at the bottom of a quartz-calcite column if the density of the column is more than 2.24 gm/cm³. In this case, where Mr. Verhoogen considers the same general phenomenon as I, some agreement is obtained. As soon as Mr. Verhoogen closes his system, however, thus abandoning the problem under discussion, his obvious conclusion is that wollastonite will not form at great depths because CO_2 is jailed in the artificial impermeable box.

One would expect to find some statements about temperature and height of the column which Mr. Verhoogen has in mind. One would also expect to find an explanation of the appearance of the somewhat mysterious value 2.24 gm/cm³. According to thermodynamics, the requirement for stabilization of wollastonite at the bottom of the column should be that the density of the quartz-calcite rock is less than about 2.90 gm/cm³, which is the density of wollastonite.

The reaction which takes place is of the following type:

$$\overset{\text{low } T}{A} \quad \overset{T_u}{\rightleftharpoons} \quad \overset{\text{high } T}{B+C}\,, \qquad (1)$$

where A and B are solids and C is solid, liquid, or gas. If $s_B > s_A > s_C$ (where s_A, s_B, etc., are the densities of the several phases), then a thick layer of A is unstable in the gravitational field even if the temperature is below T_u. A more stable state will exist if the reaction is displaced toward B and C below a certain depth, Y, so that the dense phase, B, is formed here, while the liberated phase, C, is displaced to the top of the A layer.

At stable conditions there is a certain interrelation between the thickness, Y, of the stable A layer and the existing temperature. At temperature T_u, Y is obviously zero; and the lower the temperature, the greater the thickness Y. At the depth Y we have

$$dF_A = -S_A dT + V_A dP\,, \qquad (2)$$

$$dF_B = -S_B dT + V_B dP\,, \qquad (3)$$

$$dF_C = -S_C dT + M_C dY\,, \qquad (4)$$

where F_A, F_B, etc., are the molal free energies of the several phases; S_A, S_B, etc., are the entropies of the phases; V_A and V_B are the molal volumes of the phases A and B; M_C is the molecular weight of C; and P is pressure at depth Y.

The conditions for equilibrium between A and B at depth Y and C on the top of the A layer are

$$dF_A = dF_B + dF_C\,. \qquad (5)$$

Introducing $dP = s_A dY$, and $S_B + S_C - S_A = Q/T_u$ in equation (5) gives

$$\frac{dT}{dY} = \frac{V_B s_A + M_C - M_A}{Q} \cdot T_u\,, \qquad (6)$$

which is the relation between temperature, T, and thickness, Y, of the stable A layer situated between B and C.

The reaction $A \longrightarrow B + C$ is endothermic, so that Q is positive. The requirement for gravitative splitting-up of A at temperatures below T_u is that dT/dY be negative, i.e., that $(V_B s_A + M_C) < M_A$. This condition is satisfied if $s_A < s_B$, since in this case $s_A V_B < M_B$ and $M_B + M_C$ certainly always equals M_A.

If the liberated C phase remains at the depth Y instead of being elevated, the term $M_C dY$ in equation (4) changes into $V_C dP$ or $V_C s_A dY$, and equation (6) is identical with Clapeyron's equation. It follows that if $s_A > s_C$, then $V_C s_A dY > M_C dY$, and the free energy of the system will be less if C is elevated to the top than if it remains at depth Y.

Consequently, as stated in my article, the requirement for gravitative chemical differentiation in the system $A \rightleftarrows B + C$ is that $s_B > s_A > s_C$, combined with a minimum thickness of the original A layer.

Let us take dehydration of gypsum as a simple example:

$$CaSO_4 \cdot 2H_2O \overset{T_u}{\rightleftarrows} CaSO_4 + 2H_2O\,.$$

Heat of dehydration is about 4.7 cal. and $s_{gyp} = 2.96$ gm/cm³, $s_{anh} = 2.32$ gm/cm³, $T_u =$

436°,[2] which can be introduced into equation (6); hence

$$dT/dY \cong -0.007 \text{ degrees/meters}.$$

For example, at 366° K a gypsum layer in the uppermost parts of the earth is unstable, provided that it is more than 10 km. thick. Below this depth anhydrite will tend to form, and the liberated water molecules tend to migrate toward the top. At a depth of about 30 km. the same process tends to take place at as low a temperature as 226° K.

Let us give briefly the reaction between muscovite and phlogopite in which garnet and potash feldspar are formed:

$$\underset{\text{phlogopite}}{KMg_3AlSi_3O_{10}(OH)_2} + \overset{\text{low } P\text{–}T}{\underset{\text{muscovite}}{KAl_3Si_3O_{10}(OH)_2}} + \underset{\text{quartz}}{3SiO_2} \rightleftarrows \underset{\text{pyrope}}{Mg_3Al_2Si_3O_{12}} + \overset{\text{high } P\text{–}T}{\underset{\text{potash feldspar}}{2KAlSi_3O_8}} + 2H_2O\,. \quad (7)$$

The left side corresponds to A of equation (1), pyrope to B, and potash feldspar plus water to C. The density of pyrope is 3.51 gm/cm³, that of muscovite about 2.8 gm/cm³, and that of phlogopite also about 2.8 gm/cm³.

In this case the quantity $V_{B}s_{A} + M_C - M_A$ equals about −82 gm, whereas the same quantity in the case of gypsum is about −30 gm. If the heat of transformation were known, it would be possible, at every temperature below T_u to determine the depth at which pyrope tends to form at the same time that potash feldspar and water tend to gather on the top of an idealized thick muscovite-phlogopite layer.

Mr. Verhoogen's emphatic statement that the relation between chemical potential and vapor pressure is constant by definition, regardless of the position in the gravitational field, is untenable; for the vapor pressure is independent of position in the gravitational field but the chemical potential changes according to the following formula: $d\mu/dY = M$, where M is the molecular weight of the component considered. It is obviously the "constant" K which changes in equation $\mu = RTln\pi + K$.

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[2] These data are taken from Handbook of Chemistry and Physics, ed. 30, Chem. Rubber Pub. Co., Cleveland, 1947.

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REVIEWS

"The Geological Approach to the Study of Soils." By W. H. Bryan. (Presidential Address: Section C, Geology.) (*Report of the 25th Meeting of the Australian and New Zealand Association for the Advancement of Science,* 1946.) Adelaide, 1948. Pp. 52–69.

Professor Bryan's thesis is that soil study is a necessary part of a geologist's equipment; for not only does the terrain affect the soil, but soils are themselves important geological units either as markers of epochs in stratigraphical geology or as controls of geomorphic processes.

The heads under which the thesis is developed are "Soils as Seen from Below," "Soils and Past Climates," "Influence of Landscape on Development of Soils," "Influence of Soils on Development of Landscape," and "Soils as Stratigraphical Aids."

Bryan agrees that "extensive superficial deposits may be regarded as geological formations and used in a strictly stratigraphical sense," a dictum based on Woolnough's claim that the "duricrust" of Australia—a relict layer originating from a widely developed ancient soil truncated by wind—must rank as a "stratigraphical unit"; and he stresses also the value for Pleistocene subdivision of the alternation of soils with loess formations in Nebraska and elsewhere.

Under the head "Influence of Landscape on Development of Soils" he makes the point that

> study of inherited soils found on residuals of earlier landscapes [might help] to answer the vexed question as to which, if any, of the higher remnants are truly residual surfaces. . . . All those flat-topped areas that carry a well-matured, deep soil may be confidently assigned to the original surface. . . . By combining the disciplines of geomorphology and pedology and using only those residuals that satisfy the criteria of both, there may be achieved a . . . reconstruction of the original peneplain surface. . . . Where the physiographic evidence suggests that there are present the remnants of two peneplains in the same area, a comparison of the soils of the two groups of residuals may be used to test the suggestion. . . . Yet another step . . . may be taken by the use of soils in the recognition and correlation of coeval peneplains in widely separated areas.

Examination of the "Influence of Soils on the Development of Landscape" leads to the "startling" suggestion that in the course of the normal geographic cycle, the development of characteristic and contrasting types of landscape profile is strongly influenced by soil type; for "some soils in their natural state are almost proof against erosion," while others not only facilitate runoff and sheet erosion but also are subject themselves to mass movement, soil creep, and solifluction.

> Let us examine the probable effects of two widely different soil types on the subsequent cycles of erosion. A suitable choice for our purpose would seem to be a strongly pedalferric soil such as a lateritic red earth and, for contrast, a typically pedocalcic soil such as a black earth. . . . The physical properties of such red earths as occur on small plateau remnants in southeastern Queensland are of a striking kind. The soils consist largely of an assemblage of clay minerals of low silica-alumina ratio, and consequently are remarkably friable and show little expansion on wetting. Further, the aggregation of the clay particles is such that the soil is very porous, with an internal drainage almost as free as that of loose sand. Such friable, porous clay soils are remarkably resistant to erosion—indeed they are almost erosion-proof. . . . Such soils are so absorbent that even heavy rainfall produces very little surface run-off. . . . This combination of properties provides a very effective protection for the rock beneath. . . . In such circumstances the erosive forces acting upon our uplifted peneplain will be concentrated in and restricted to a small number of very widely spaced streams fed almost entirely by ground water, in each of which vertical corrasion will be far more effective than lateral erosion. The characteristic plateau topography thus developed early in the cycle will be reflected right through to the end. . . . Residuals . . . will remain steep-sided, flat-topped, and soil-protected to the end. . . .
>
> In vivid contrast to such a picture is that to be deduced when the original soil cover of the uplifted peneplain is of a strongly pedocalcic nature such as a black earth. . . . This black soil too consists largely of clay, but this clay is made up of minerals of a high silica-alumina ratio, and is notoriously adhesive and tenacious. It expands greatly on wetting, is very easily eroded, and, on even moderate slopes, is very subject to landslides. The effect of rain is quickly to saturate the upper part of the soil, thus forming a sodden, impervious mass over which the excess water flows . . . carrying with it considerable quantities of soil, the result of natural sheet erosion. The civil en-

gineer detests these black soils, which are as dynamic as the red earths are static....

The picture that we now observe is in sharp contrast with that due to the protecting influence of the red earths. Far from acting as a protective cover of the original surface, the black earths will tend to move laterally into valley depressions as soon as the valley sides are sufficiently inclined to permit movement down their slopes. As a result of the continuous migration of soil from the intervening ridges the rocks beneath these are recurrently exposed to provide more soil, further to feed the valleys and thus perpetuate the original landscape pattern at progressively lower levels. From the very beginning the landscape takes on an "old" appearance, which it retains through youth and maturity until it becomes old in fact as well as in appearance. Throughout the whole cycle the "rolling downs" motif is manifest. "Black-soil plains," as they are called, are similar in many parts of the world, and all conform to one geomorphic type in spite of the fact that they must represent different stages in the cycle of erosion. Consequently such areas give no clue as to the position of the original peneplain from which their present surfaces were derived. Whereas the landscape developed under the red earths jealously preserves the proof of its origin, that covered by the black earths automatically effaces the evidences of its own history.

C. A. COTTON

Sequence in Layered Rocks. By ROBERT R. SHROCK. New York: McGraw-Hill Book Co., Inc., 1948. Pp. 507; figs. 397. $7.50.

The author's purpose is to "describe and illustrate top and bottom features in all kinds of bedded and layered rocks." It is surprising, at first thought, that this topic could be the subject of a 507-page book. A wide range of material has been consolidated into an almost complete compendium of what is known about one of the most fundamental concepts of geology—superposition.

The book is descriptive and extremely well done. A total of 397 sketches and photographs evidences the care taken in the preparation of topics. Abundant quotations from published papers and numerous personal communications are included. Complete references to quoted material facilitate further inquiry into a particular subject. It is regrettable that the extensive non-English literature has not been consulted in the preparation of the book.

The completeness of the work would have been tremendously enhanced if the author had included such allied subjects as sedimentation rates, the problem of geologic time, evaluation of geophysical well logging in determining sequence, and a fundamental discussion of time, lithologic, and fauna surfaces.

N. A. R.

Chambers's Mineralogical Dictionary. Brooklyn: Chemical Publishing, Inc., 1948. Pp. 47; colored pls. 40. $4.75.

This booklet consists primarily of an alphabetical listing of minerals, rocks, and structures. Some fourteen hundred items appear in fine print. The definitions given are popular and of no value to the serious student of the subject. The only illustrations are very mediocre color plates. These are indexed.

D. J. F.

ERRATA

In "The Unique Association of Thallium and Rubidium in Minerals," by L. H. Ahrens, in the November, 1948, *Journal:*

P. 578, col. 1, line 13, *read* PbS *instead of* Pbs.

P. 579, col. 1, line 20, *read* Cs^+ *instead of* Cs.

P. 580, col. 2, line 3, *read* Rb_2O/Tl_2O *instead of* RB_2O/Tl_2O.

P. 585, col. 1, line 6, read $x = 0$ instead of $x = O$.

P. 587, col. 2, line 1, *read* "alkali-metal" *instead of* alkali-metal.

P. 588, table 4, col. 3, line 10, *read* 0.00031 *instead of* 0.0031.

P. 590, col. 1, line 6, *read* "alkali-metal" *instead of* alkali-metal.

COMMUNICATIONS AND ANNOUNCEMENTS

NATIONAL RESEARCH COUNCIL COMMITTEE ON THE INTERRELATIONS OF PLEISTOCENE RESEARCH

A broad understanding of earth conditions during the Pleistocene has become necessary in fields of scientific endeavor other than those of geology and prehistoric archeology. In studies as diverse as pedology, climatology, biology, and oceanography, productive research effort requires considerable knowledge of progress made in the related fields of Pleistocene investigation. As a means of promoting fruitful contacts among workers in these several sciences, the National Research Council established in June, 1947, a Committee on the Interrelations of Pleistocene Research under the chairmanship of Professor Richard Foster Flint, of Yale University. Working with the chairman are the following: Edward S. Deevy, Jr. (biogeography), Department of Biology, Yale University; Loren D. Eisley (anthropology and New World archeology), Department of Anthropology, University of Pennsylvania; Claude W. Hibbard (vertebrate paleontology), Museum of Paleontology, University of Michigan; Chauncey D. Holmes (glacial erosion and sedimentation), Department of Geology, University of Missouri; Helmut E. Landsberg (meteorology and climatology), Committee on Geophysical Sciences, Research and Development Board, Washington, D.C.; Hallam L. Movius, Jr. (Old World archeology), Peabody Museum of Archeology and Ethnology, Harvard University; Fred B. Phleger, Jr. (oceanography and sea-floor geology), Department of Geology, Amherst College; Louis L. Ray (glacial stratigraphy; alpine glacial geology), United States Geological Survey, Washington, D.C.; H. T. U. Smith (eolian features; frozen ground; stream terraces), Department of Geology, University of Kansas; and James Thorp (soil science), Division of Soil Survey, United States Department of Agriculture, Lincoln, Nebraska.

Preliminary reports were circulated among the membership, and on April 5 and 6 the committee held its first meeting. The results are most encouraging. To make its work as effective as possible, the committee welcomes correspondence with scientists in any field in which the combined resources of the committee can be of assistance. Such correspondence should be directed to the individual member representing the particular field, or to the chairman.

CHAUNCEY D. HOLMES

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SYMPOSIUM ON MINERAL RESOURCES

The Department of Geology and Geography of the University of Tennessee is sponsoring a symposium on the mineral resources of the Southeast. Sessions will be held on the campus at Knoxville on March 3, 4, and 5. Papers on the major metallic and nonmetallic mineral products of the Southeast will be presented by invited speakers.

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THE JOURNAL OF GEOLOGY

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THE GEOLOGIC IMPLICATIONS OF CRYOPEDOLOGY[1]

KIRK BRYAN
Harvard University

GENERAL STATEMENT

The present issue of the *Journal of Geology* presents two classes of papers: those recording studies made in the arctic and those describing and interpreting phenomena in areas of temperate climate formed by processes now current in arctic areas. Thus the papers present two aspects of cryopedology or the study of the effect of frozen ground and intensive frost-action.

It is gratifying to be asked by the Editor of the *Journal* to write this introduction, but it is still more pleasant to recall that I have been in the confidence of many of the authors and have spent with them happy hours in the field and in the laboratory.

THE STUDY OF INTENSIVE FROST-ACTION

The breakup of rocks by the freezing and consequent expansion of water in cracks is commonplace in the temperate climate of northwestern Europe and northeastern North America. Our geologic textbooks devote a paragraph or two to the process. Engineers recognize that resistence to the process is a necessity for good building stone and that frost is deleterious to concrete. In the arctic and in high mountains the process is much accelerated and for practical purposes the breakup of rocks is almost entirely by congelifraction However, we lack detailed knowledge, although there seems to be some reason for believing that the silt size is the lower limit of comminution of rock. This broad generalization has not yet been confirmed by adequate observation, and we have as yet little information on the rate of rock destruction.

The prime feature of the arctic is the existence of perennially frozen ground ("pergelisol"). The character, thickness, and extent of pergelisol are still the subject of study. It is, however, evident, as shown by studies in the region of Petsamo (Aario, 1943) and also by some of the present papers, that this phenomenon was more extensive in the past than it is at present. Further, the area of pergelisol is greater at present than it was in the postglacial warm period. Thus the formation of pergelisol, or "pergelation," and the decay of pergelisol, or "de-pergelation," are of prime importance in any study of arctic phenomena.

The existence of frozen ground largely prevents infiltration of water into the

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[1] Manuscript received January 28, 1949.

underground, and hence water at the surface is ubiquitous in the melting season. Thus the arctic, in spite of a low precipitation and relatively large evaporation, has in summer much wet and soggy ground. The zone of annual melting, the "mollisol" or active layer, is the zone in which movement of the surface material occurs. The character of this movement appears to depend on the available water, the grain size, the surface gradient, and the vegetation. Studies of present-day "soil structures" and movements will define these interrelationships. Further, the study of intensifications in the past, such as those here described by Wahrhaftig, provide a comparison leading to interpretation of the severity and duration of the colder climates of the past within the unglaciated portions of the arctic.

If, even in the arctic, fluctuations in the area and thickness of perennially frozen ground have occurred, one wonders whether the numerous lakes and ponds which appear to be due to melting of the pergelisol are the result of local conditions, such as the growth and more or less accidental destruction of the vegetation cover, or to de-pergelation. It seems probable that the well-known retreat of Alaskan glaciers is due to a recent amelioration of climate. If so, this amelioration may at the moment be inducing a state of de-pergelation that affects present-day surface features and induces the formation of lake basins by melting of ice in the ground.

THE PERIGLACIAL CLIMATE

That certain surface deposits consisting of rubbles of various types could not have reached their present place by any process now in action was recognized more than one hundred years ago. In England, Godwin-Austin (1840) and Fisher (1866) correctly attributed these materials to intensive frost-action in the past. In 1896 Blankenhorn, on the basis of English studies, attacked the same problem in Germany. But the main outburst of interest followed the excursion of the International Geological Congress to Spitzenbergen in 1910, led by B. Högbom. Högbom's great paper of 1914 and numerous papers by German authors of 1912 and in following years by Europeans in general have built a great accumulation of data on the character and distribution of deposits in temperate Europe attributed to intensive frost-action in the past.

Lozinski (1909) introduced the term "periglacial" for (1) the area adjacent to the border of the Pleistocene ice sheets; (2) the climate characteristic of this area; and (3), by extension, the phenomena induced by this climate even if located outside the main periglacial zone. Lozinski realized that, however low the mean temperature at the ice front, the climate in the latitude of Germany or any other now temperate region could not be truly arctic because of the difference in position of the sun. Troll (1947) suggests that the zone of perennial snow be called the "nival climate" and that the zone of intensive frost-action be called the "subgelid climate" ("below the ice" or firn-line). In the arctic and presumably during the ice advances of the Pleistocene in temperate regions, the zone of frost-action is marked by pergelisol. But Troll holds that in many mountain areas there is no pergelisol, only diurnally or annually frozen ground.

The position of the lower limit of the subgelid climate conforms to that of the lower limit of the snow-line. It rises along a meridian from sea-level to a high point in the subtropic arid belt and falls in elevation near the equator. It rises inland from the western side of continents and has many irregularities in elevation by reason of orographic relations. Furthermore, the intensity of frost-action is affected not only by climate but by edaphic relations.

Whether it is worth while to distinguish the climate with the name "subgelid" is questionable. After all, during the Pleistocene, the large part of the area involved lay south of and bordering the ice sheets and was literally periglacial. In mountain regions of all latitudes the zone lies below the snow-line or firn-line and hence, from an areal standpoint, surrounds the mountain glaciers and is thus periglacial.

The amount and intensity of frost-action depends on climatic factors, and in the world-wide lowering of temperature during the Pleistocene the periglacial zone was much expanded.

The winter season may have had continuous temperatures below freezing but was probably not so long as the present arctic winter. The fall and spring transitions were probably marked by more numerous fluctuations from cold to warm, i.e., "crossings of the frost line." The summer season probably had warmer periods than any characteristic of the present arctic summer but may easily have had cold periods much below freezing. Thus frost-action may have been much more intense in the periglacial zone than in most of the present arctic. Renewed study of the arctic will presumably define the possibilities and will make much surer our interpretation of the character of the periglacial zone.

The extent of the periglacial climate is one of the problems of the future. As Troll (1947) points out, the world-wide lowering of snow-line during each cold stage of the Pleistocene must have produced a lowering similar, but not necessarily equal, to the lowering of the zone of intensive frost-action. We must then consider the amount of modification of the Tertiary topography by frost-action, i.e., by cryoplanation. How great is this modification? To what extent does recurrent cryoplanation during the Pleistocene invalidate some of our previous analyses of land forms in areas south of the continental ice sheets and in mountainous areas whose higher peaks bore glaciers during the Pleistocene?

Studies of ancient frost-action in temperate areas and of the wind systems of the Pleistocene in similar areas, such as are published in this series of papers, furnish data on the periglacial climate. They mark a step forward in the definition of its area and intensity. Thus they furnish a beginning on which a more complete interpretation of the periglacial climate can eventually be made. With this interpretation, some of our general geomorphological problems will be closer to solution. The present interest in the development of mountain and hill slopes promises to fill a gap in geomorphologic analysis. We must, however, be sure that in our study of slopes the fluctuations of Pleistocene climate are considered. Are the present-day processes solely responsible for existing slopes? Or are existing slopes residual from the processes of a past periglacial climate?

Analysis of the processes of weathering is also involved. As shown herewith by Schafer, the present soil in central Montana has developed on a mass of material contorted by congeliturbation. Part of this mass is an ancient soil. In places the present soil is developed on materials of the ancient soil and is therefore polygenetic. The recognition of the polygenesis of many soils is proceeding rapidly (Nikiforoff, 1948). In this field of effort, studies similar to those set forth in this group of papers have a very large importance. The detailed studies of soil scientists will also add a precision that should be welcome to geomorphologists.

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ORIENTED LAKES OF NORTHERN ALASKA[1]

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ABSTRACT

The oriented lakes of northern Alaska occur in an area of more than 25,000 square miles in the Arctic Coastal Plain Province. The topography, drainage, vegetation, climate, geology, and permafrost of the area are briefly described. The average range of orientation of the lakes is 12°—from N. 9° W. to N. 21° W. In any locality the deviation from the average is commonly less than 3° and rarely over 5°. The lakes range in size from small ponds, a few tens of feet in length, to large lakes more than 9 miles long and 3 miles wide. The shapes may be described as elliptical, cigar-shaped, rectangular, ovoid, triangular, irregular, or compound. One group of lakes has a shallow shelf or underwater bench surrounding a deeper central portion. The rest are shallow throughout, and the underwater profile is commonly concave. The major outline of the lakes is smoothly curved, but in detail it is cuspate or jagged. No lacustrine beach ridges were recognized. Former lake basins, now drained, and extensions of the present lakes are evidenced by shore features, lacustrine deposits, the character of polygonal ground, and vegetation. The effect of wind, vegetation, and permafrost are briefly described. The lakes are compared with the Carolina Bays and with rectangular lakes in eastern U.S.S.R. Many of the lakes are believed to be the result of thawing of permafrost; others may be produced by the segmentation of uplifted lagoons. The origin of some is not known.

INTRODUCTION

GENERAL STATEMENT

The oriented lakes of northern Alaska occur in a remote and sparsely inhabited area of more than 25,000 square miles adjacent to the Arctic Ocean (fig. 1 and pl. 1, *A*). The origin of the thousands of lakes has aroused the curiosity of most men traveling on the Arctic slope; however, few systematic data have been collected regarding their origin or their remarkable parallelism. The areal extent and orientation of the lakes probably has been recognized only since observations could be made by air or from aerial photographs.

METHOD OF INVESTIGATION

The oriented lakes first came to the attention of the senior author, R. F. Black, during the summer of 1945. In the summer of 1946, in connection with the permafrost investigations of the United States Geological Survey, W. L. Barksdale visited numerous lakes along the Meade River and in the vicinity of Barrow. R. F. Black, assisted by Donald R. Loftus, visited many lakes along the Arctic coast southwest and southeast of Barrow, in the vicinity of Barrow, and in the Cape Simpson area. Several trips were made by air from Barrow south and east over the lake country.[2] Trimetrogon photographs, taken in 1942, 1943, and 1945 of most of northern Alaska, were used as a base for delimiting the areal extent of the lakes and for providing data in places not visited. Planimetric maps, on the scale of 1/48,000, made by the Geological Survey from these aerial photographs, provide coverage for most of the entire area.

The authors have not had an opportunity to study in detail the many aspects of the oriented lakes; however, in

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[2] In 1947, after this paper had been written, R. F. Black made numerous trips by float plane from Barrow and Umiat to lakes in the Coastal Plain west of the Colville River. The southern margin of the Coastal Plain was traversed, and much additional data on the distribution of soils, vegetation, oriented lakes, and geology were accumulated. They serve to substantiate the findings herein reported, but are not incorporated into this paper.

the light of the increasing interest in Alaska it seems desirable to present the available information at this time. The Geological Survey anticipates further studies of the oriented lakes and associated permafrost phenomena.

DESCRIPTION OF AREA

Philip S. Smith and J. B. Mertie, Jr. (1930), have summarized the geography and geology of northwestern Alaska and have listed publications which contain complete bibliographies and accounts of previous explorations. Recent unpublished studies by the Geological Survey in and adjacent to Naval Petroleum Reserve No. 4 have modified their conclusions and added considerable new information.

TOPOGRAPHY

The area occupied by the oriented lakes of northern Alaska covers almost completely the physiographic province known as the Arctic Coastal Plain (fig. 1). The Coastal Plain is the northernmost physiographic province in Alaska and is bordered on the south by the Arctic Plateaus Province. The boundary between the Coastal Plain and the Plateaus has been arbitrarily chosen on the basis of topography and of distribution of Quaternary sediments (Smith and Mertie, 1930). In general, that boundary is between 400 and 600 feet in altitude, although Leffingwell (1919, p. 53) reports that on the Okpilak River, 40 miles from the coast, the boundary is about 1,200 feet. East of the Sadlerochit River (outside the area of the oriented lakes and not shown on fig. 1) the Coastal Plain widens abruptly to about 50 miles (Leffingwell, 1919, p. 53) and narrows where the British Mountains approach the ocean at the international boundary line.

In some places the Coastal Plain grades imperceptibly into the northern front of the Arctic Plateaus. In the Canning River area (Leffingwell, 1919, p. 52) the northern front of the Arctic Plateaus rises with a slope of 15°–20° from the flat Coastal Plain and may be as much as 300 feet higher.[3]

To the north the profile of the Coastal Plain is broken only slightly by the Arctic Ocean. In many places the plain is only 1–10 feet above the sea. The average height of coastal bluffs is only 15 feet, and the highest rise 25–50 feet above sea level. The shoreline is characterized by many offshore bars of silt and sand, mud flats, and spits. Lagoons, in places as much as 5 miles wide and many tens of miles long, are generally less than 10 feet deep.

Except for minor features, the Coastal Plain extends with monotonous flatness from horizon to horizon. Broad low ridges and valleys have a very low relief of about 30 feet. Locally steeper slopes of former wave-cut benches or wave-built benches or beaches occur. Small earth and ice mounds, commonly 1–4 feet high, occur over most of the plain. Shallow troughs, commonly less than 2 feet deep and in many places underlain by vertical wedges of ice, form a network (polygonal ground) over most of the surface.

DRAINAGE

All runoff is on the surface, as permafrost underlies the entire Coastal Plain at a depth of a few inches. All but the largest rivers, such as the Colville River, freeze solidly to permafrost every winter. It is estimated that possibly 50–75 per cent of the Coastal Plain is covered with lakes and marshy ponds. Many are without external drainage except in spring,

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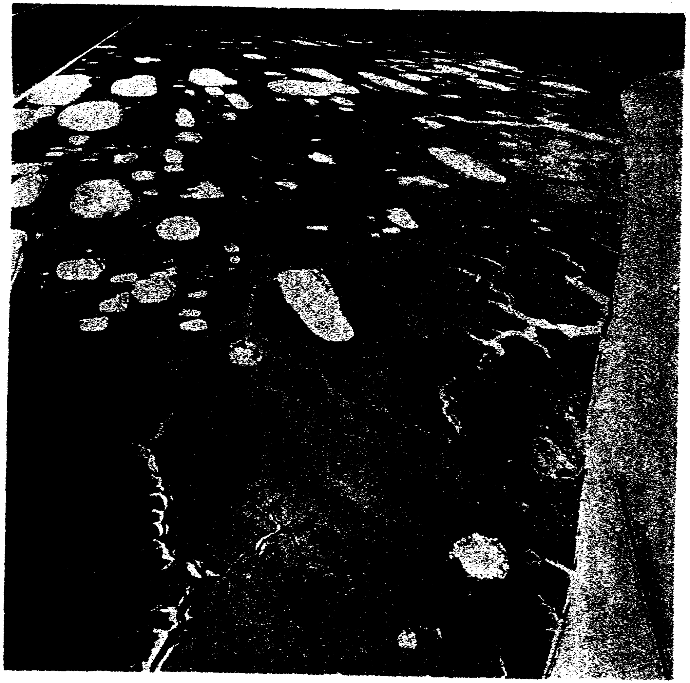
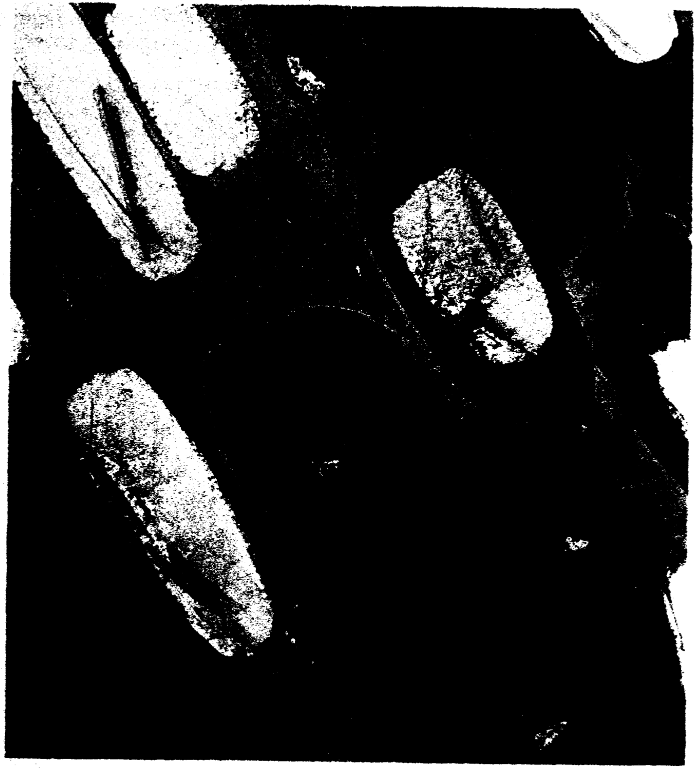
[3] R. F. Black (1947) found that, except for minor breaks, the boundary between Cape Beaufort and the Colville River is a distinct topographic break, as much as 300 feet high in places but generally lower.

A, Oblique aerial photograph of oriented lakes on the Arctic Coastal Plain south of Barrow, Alaska.

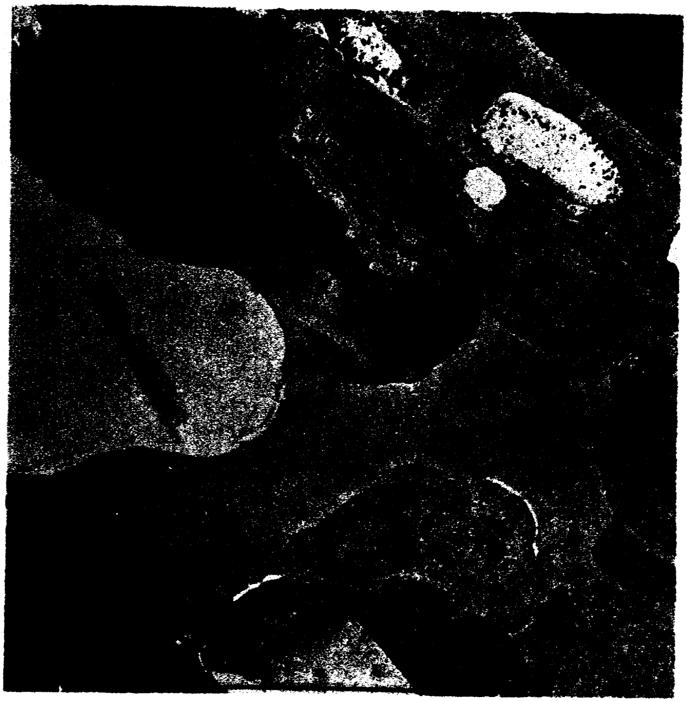


B, Oblique aerial photograph of oriented lakes on Coastal Plain south of Wainwright, Alaska.

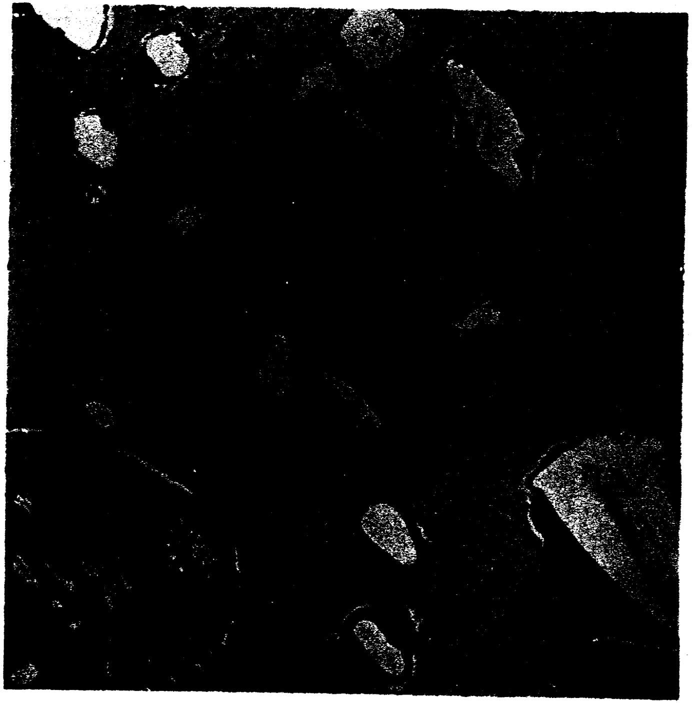


A, Vertical aerial photograph of oriented lakes with relatively smooth shorelines.

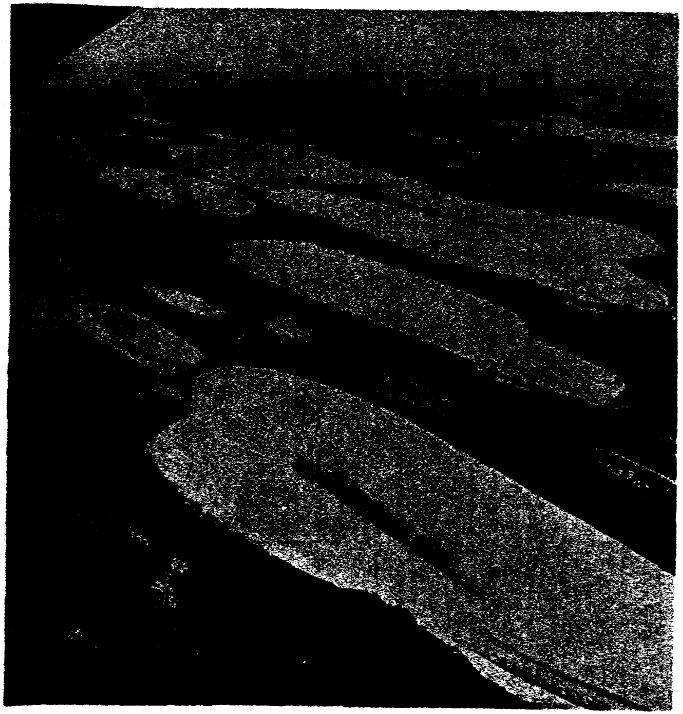
B, Vertical aerial photograph of oriented lakes with cuspate or jagged shorelines and shallow underwater "benches."



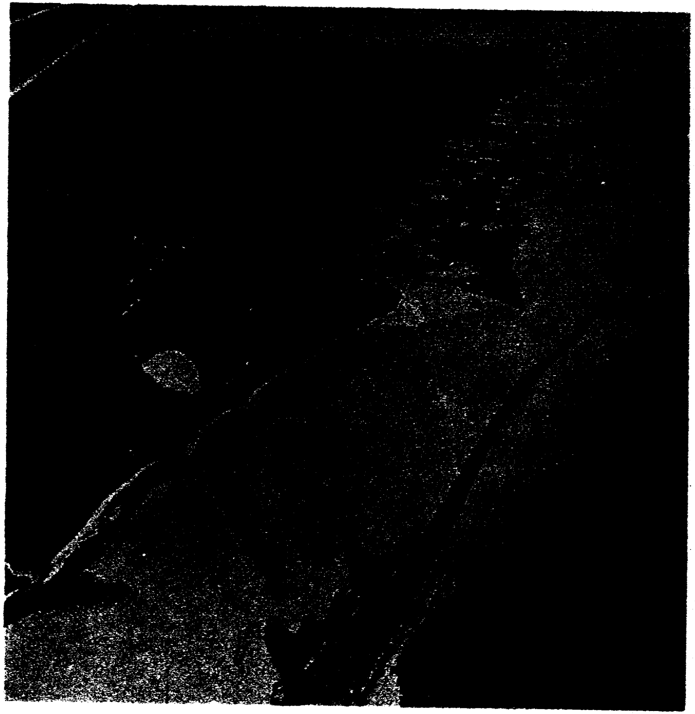
A, Vertical aerial photograph of three generations of lakes.



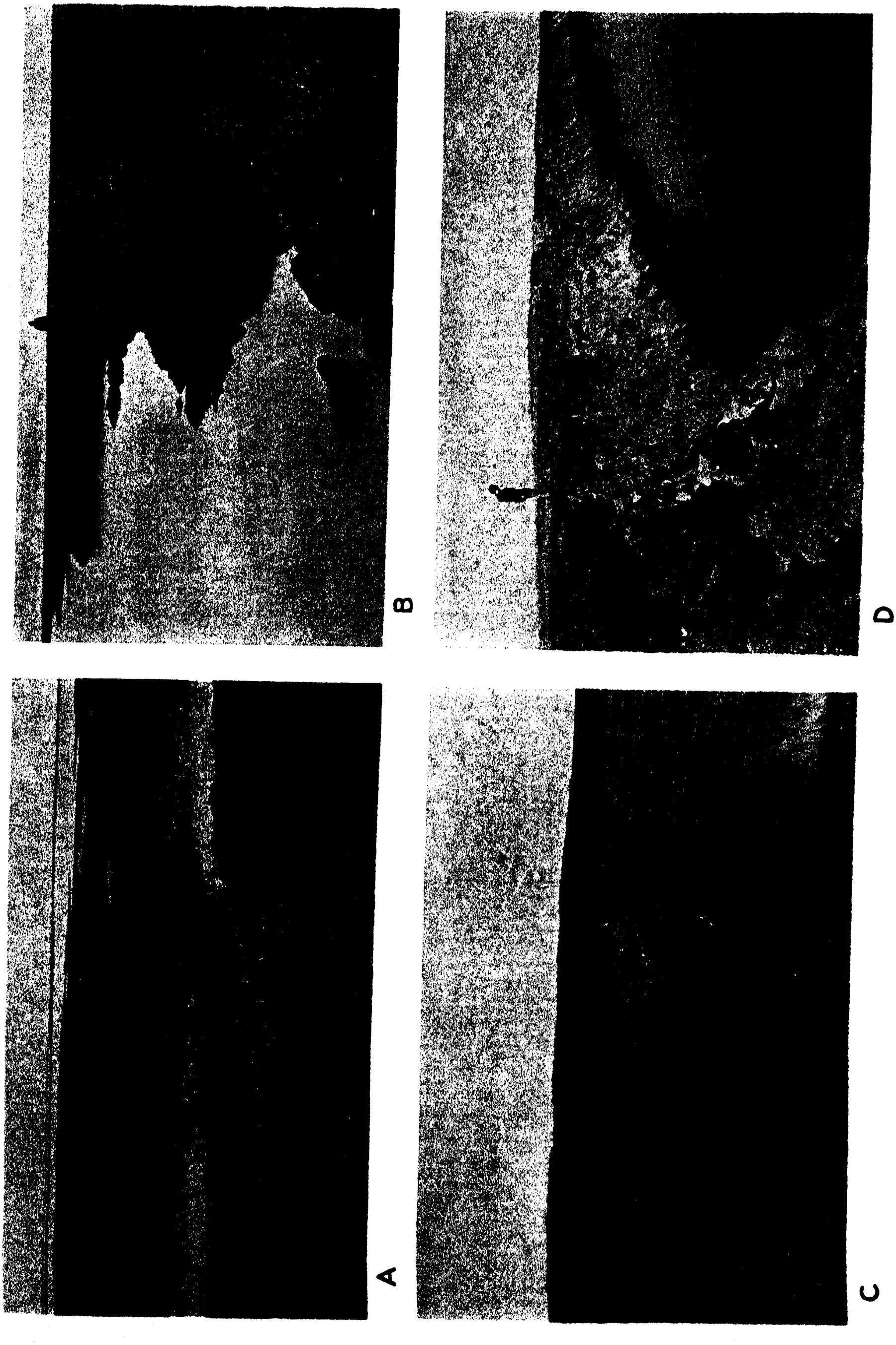
B, Vertical aerial photograph of sand dunes and oriented lakes, showing effect of wind.



A, Oblique aerial photograph of coalescing lakes southeast of Barrow.



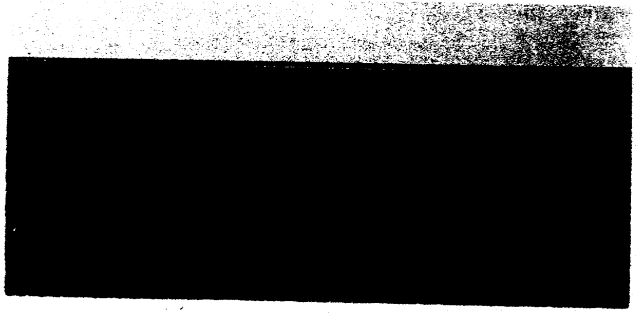
B, Oblique aerial photograph of the Coastal Plain near Peard Bay, showing part of an uplifted and segmented lagoon.



Shore features of oriented lakes

A

A. Beach and peaty silt bank containing ice wedges on the east shore of the lake one-half mile southeast of the Barrow airbase.

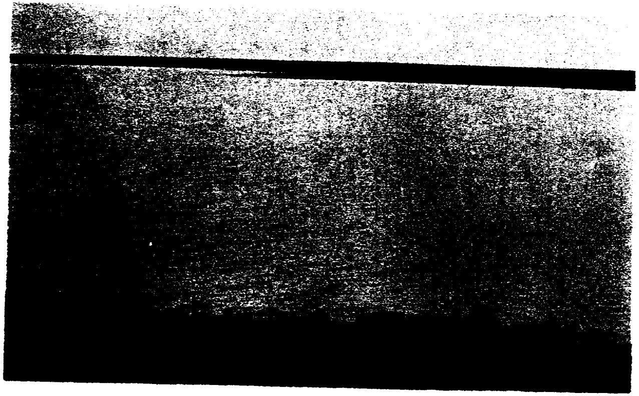


B

B. Relatively well drained tundra in the southern part of the Coastal Plain, showing the incised character of an oriented lake one-quarter mile distant.

C

C. Oblique aerial photograph of oriented rectangular lakes in the valley of the Penzhina River, eastern U.S.S.R. Compare with pls. 1, 2, 3, and 4, *B*.



D

D. Oriented lake shown in *B*.

when high waters overflow their banks. Certain localities are partially dissected, and some lakes are draining through recent gullies.

Only one large river, the Inaru, lies entirely within the limits of the Arctic Coastal Plain Province. However, most rivers that head in the Arctic Plateaus Province or on the north flanks of the Brooks Range cross the Coastal Plain on their way to the Arctic Ocean.

VEGETATION

The entire Coastal Plain is a treeless, prairie-like tundra region characteristic of the arctic and subarctic zones (Palmer and Rouse, 1945, p. 1). It is covered with a thick mat of vegetation composed of a mixture of lichens, mosses, grasses, sedges, shrubs, and other plants. Small prostrate shrubs are rare along the coast but are more abundant inland. Willows are particularly common along the protected banks of the larger streams, though they rarely grow as high as 10 feet. Evergreen or large deciduous trees are absent.

CLIMATE

The climate on the Arctic slope is cloudy, cold, arid, and windy. The only weather station[4] on the northern Coastal Plain of Alaska is at Barrow, but data collected there since 1912 by the United States Weather Bureau (1943) are probably applicable to the entire coastal area. Monthly and annual climatic data are given in table 1.

TABLE 1

MONTHLY AND ANNUAL CLIMATIC DATA AT BARROW, ALASKA

(U.S. Weather Bureau, 1943)

	Av. No. Days with Freezing Temperatures	Av. Wind Velocities (m.p.h.)	Mean Precipitation (Inches)	Snowfall (Inches)	Cloud Cover (Per Cent)
Jan.	31	10.0	0.15	2.6	42
Feb.	28	9.9	0.20	2.7	24
Mar.	31	11.2	0.13	2.0	31
Apr.	30	11.9	0.12	2.7	45
May	30	11.4	0.14	1.8	71
June	23	11.0	0.26	0.4	61
July	13	12.5	0.89	1.0	62
Aug.	16	12.9	0.73	0.7	82
Sept.	27	14.1	0.49	3.0	82
Oct.	31	15.0	0.56	7.9	72
Nov.	30	11.7	0.30	4.3	46
Dec.	31	10.4	0.26	3.9	48
Annual	321 (21–23-year record)	11.8 (10-year record)	4.23 (25–27 year record)	33.0 (21–23-year record)	56

The annual mean temperature at Barrow is 10° F.; the annual mean maximum temperature is 16° F.; and the annual mean minimum temperature is 4° F. The monthly mean maximum and mean minimum temperatures are shown in figure 2. The absolute maximum temperature is 78° F. and the absolute minimum temperature is $-56°$ F. The monthly and annual number of days with freezing temperatures are shown in table 1.

During a period of 27–30 years the most frequent wind direction was northeast during every month except July, when it was southwest. The prevailing direction and relative velocity of the wind at Barrow are shown in figure 3. The average monthly and annual wind velocities are shown in table 1.

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[4] Since this paper was written, a weather station has been started at Umiat, on the Colville River.

GEOLOGY

The Arctic Coastal Plain is underlain by several thousand feet of relatively impervious sandstones, shales, and some interbedded conglomerates, coals, limestones, and bentonites of Upper Cretaceous and early Tertiary age. These rocks are essentially flat-lying, although gentle folding has produced large low domes and anticlines and broad, shallow basins and synclines. Dips are commonly less than 5°.

The Upper Cretaceous and early Tertiary rocks are mantled uniformly throughout much of the Coastal Plain by unconsolidated clays, silts, sand, gravel, and peat, as much as 150 feet thick, of late Tertiary-Quaternary age. The unconsolidated sediments are of interfingering marine and fluvial origin, and different deposits vary widely in areal distribution, thickness, and composition. The detailed geologic history and relations of these deposits are known for only a few areas.

PERMAFROST

All surficial deposits and the bedrock to a depth of several hundred feet under the Arctic slope are permanently frozen. The surficial deposits, in particular, contain large quantities of ice in the form of nearly vertical ice wedges, horizontal ice sheets, irregular ice masses, and small grains, crystals, stringers, irregular particles, and films of ice as a cement. Shallow thawed zones probably occur under the larger streams and below some of the deeper lakes. Summer thaw penetrates from 6 inches to a maximum of about 30 inches. The average summer thaw is about 8–20 inches (Black and Barksdale, 1948; Black, 1947).

DESCRIPTION OF THE ORIENTED LAKES

DISTRIBUTION

The oriented lakes occupy almost completely the Arctic Coastal Plain Province of northern Alaska and, as far is now known, are limited to that province (fig. 1). The northern limit of the lakes is the Arctic Ocean. The southern limit coincides approximately with the boundary between the Coastal Plain and the Plateaus Province, from the Arctic Ocean near Cape Beaufort, about 450 miles eastward, to the Arctic Ocean near the Kalakturuk River. Although the Arctic Coastal Plain Province continues east of the Sadlerochit River to the international boundary, the lakes in that area are apparently not oriented. The maximum width of the oriented-lake area, from Barrow southward, is more than 100 miles. The average width is between 50 and 80 miles.

Oriented lakes occur chiefly in lowland areas and less commonly on the sides or tops of low ridges. Many are in basins cut directly into broad, featureless plains. Others are present in lowland areas along coasts, on tops of high bluffs, and less commonly in river floodplains.

CONCENTRATION

The oriented lakes are concentrated in elongated belts or zones, which apparently represent former stages in the uplift (Smith and Mertie, 1930, p. 48) of the Coastal Plain. It is estimated that locally the lakes or drained lake basins comprise from less than 25 per cent to as much as 90 per cent of the surface of the land, although in most places the lakes probably comprise between 50 and 75 per

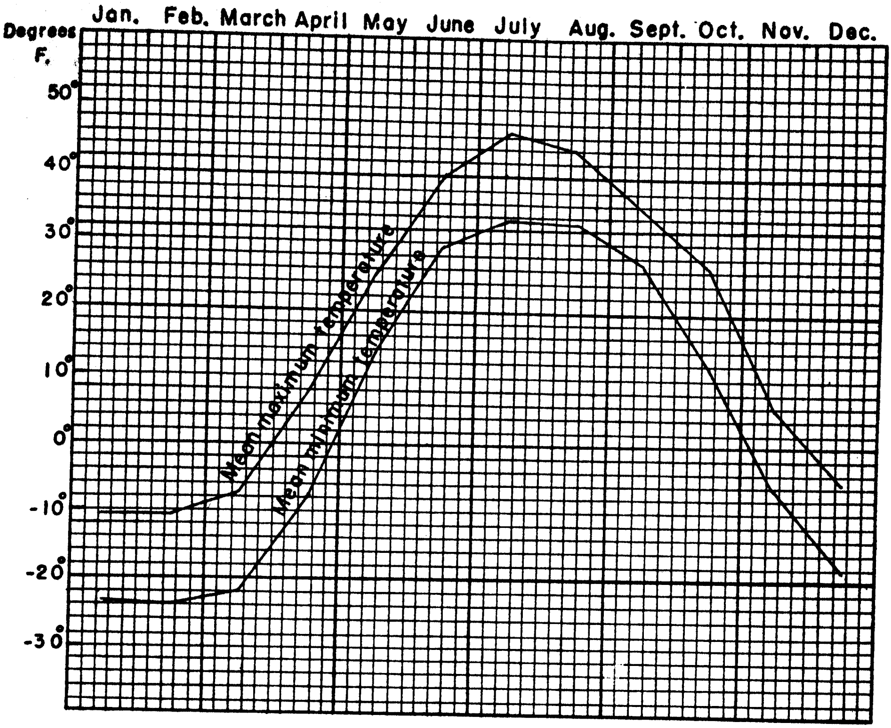


FIG. 2.—Mean maximum and minimum temperatures at Barrow, Alaska

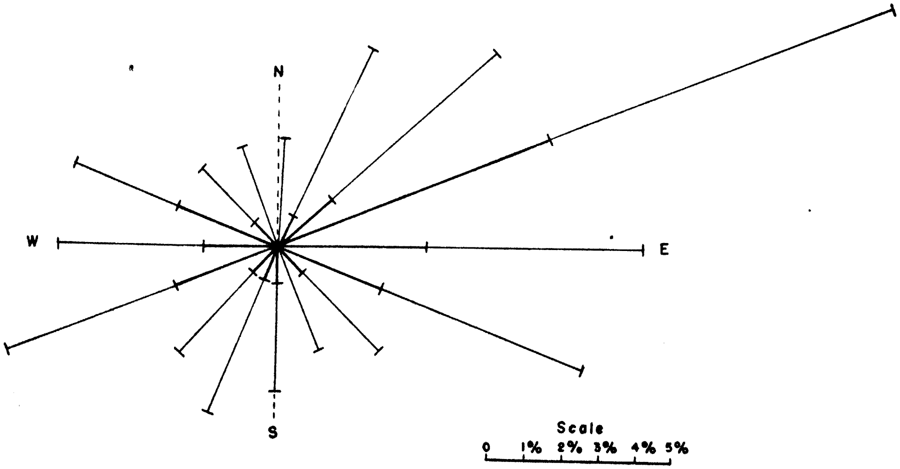


FIG. 3.—Direction, velocity, and frequency of winds at Barrow, Alaska. Length of vector indicates percentage of time that wind blows in indicated direction. Winds blow toward center. Heavy line indicates winds of velocity greater than 15 m.p.h. (31 per cent of total). Light line indicates winds of velocity from 1 to 15 m.p.h. (69 per cent of total). Calms exist less than 1 per cent of the time.

cent of the surface. Most of the oriented lakes are concentrated in a large area north of latitude 70° and between the Kuk River on the west and the Itkillik River on the east. The number of oriented lakes is markedly less along the southern boundary of the area, and particularly so in two areas of slightly higher relief and coarser-grained surficial deposits, one of which lies between the Itkillik and Kuparuk rivers and the other between the Sagavanirktok and Shaviovik rivers, where only remnants of former oriented lakes remain. In certain areas adjacent to the coast, along many of the larger rivers, and wherever slopes are moderate, gullies have intersected and largely drained the oriented lakes without, however, completely destroying the lake basins themselves.

ORIENTATION

The remarkable parallelism exhibited by the thousands of elongated lakes in the Arctic Coastal Plain Province in northern Alaska is a rare phenomenon and, so far as is known, is duplicated in such a marked degree only by the Carolina "Bays" of the Atlantic Coastal Plain Province of the United States. The uniformity of alignment of the lakes is so perfect in many parts of the Arctic Coastal Plain that their orientation has been used as an aid in air navigation (see fig. 1 and pl. 1, *A*).

In order to determine accurately the regional orientation of the lakes, the long axes of five to twenty lakes in a particular locality were determined by bisecting the outlines of the lakes as shown on Geological Survey planimetric maps; the trend of the axes was measured; and the average trend determined arithmetically. The number of lakes measured was increased as the variability of orientation increased. Of the well-developed lakes oriented in a northerly direction, the total range in orientation is only 30°—from north to N. 30° W. The average range for seventeen localities, as shown by the arrows on figure 2, is only 12°—from N. 9° W. to N. 21° W. In any one locality the deviation from the average is commonly less than 3° and rarely over 5°. The lakes in the southeastern part of the Coastal Plain in the vicinity of the Colville River at latitude 70°00′ trend more nearly west than do the lakes in the rest of the Coastal Plain. Most lakes in the central portion of the Coastal Plain north of latitude 70°00′ and between the Kuk River on the west and the Ikpikpuk River on the east trend N. 13° W. to N. 15° W.

Not all lakes on the Coastal Plain are oriented, nor are all oriented lakes aligned in the same direction. The north-northwest trend is uniformly developed throughout the region. An easterly trend is shown by a few lakes in scattered localities but is nowhere so pronounced as is the north-northwest trend. Rarely, individual lakes along the coast or in the floodplains of some of the major rivers are oriented in other directions, but no other pronounced trends were noted.

Lakes which are not oriented occur in the area along the southern limits of the Coastal Plain, in the areas of higher topographic relief between the Itkillik and Kuparuk rivers and the Sagavanirktok and Kadleroshilik rivers, in certain relatively high sections along major streams, and on the floodplains of many streams.

The oriented lakes in some places form groups or chains that are aligned along the axis or at an angle to the axis of individual lakes composing the chain.

CHARACTER OF THE ORIENTED LAKES

Size.—The oriented lakes range in size from small ponds a few tens of feet in length to large lakes more than 9 miles long and 3 miles wide (fig. 4). Lakes of all sizes occur in any locality, although large lakes are more numerous in a broad zone adjacent to the coast from Barrow eastward and as far south as latitude 70°30′. Many of these lakes are 4–6 miles long and ½–1 mile wide. Lakes are conspicuously less numerous and smaller east of the Itkillik River than west of that river, and they are, in general, less numerous and smaller toward the southern boundary of the Coastal Plain.

Lake Teshekpuk, about 80 miles southeast of Barrow, is an unusually large and irregular lake more than 21 miles wide and 29 miles long.

Shape.—The shapes of the lakes of the Coastal Plain may be described as elliptical, cigar-shaped, rectangular, ovoid or egg-shaped, triangular, irregular, or compound, having any combination of these shapes. All the lakes are somewhat irregular, and few are so smooth or so perfectly formed as to fit clearly into any one type. Any one or more types may predominate in a particular locality, although in most areas all types are at least represented.

Representative examples of the various types have been traced from Geological Survey maps and are shown in figure 4: triangular to ovoid types are shown in *A*, *G*, and *H;* rectangular to cigar-shaped types in *B*, *C*, *L*, and *O;* ovoid to elliptical types in *D;* rectangular types in *E;* compound types in *I*, *K*, *L*, and *M;* compound to irregular types in *J;* rectangular to irregular types in *N;* and a cluster of all types in *F*.

Ratio of length to width.—The ratio of length to width is as variable as the shapes of the lakes; it ranges from almost 1:1 to more than 5:1. Many well-developed elongate lakes in the central portion of the Coastal Plain, north of latitude 70°00′ and between the Kuk and Ikpikpuk rivers, have ratios of 2:1 to 3:1; and many lakes in a zone 20–50 miles wide along the coast from the Kuk River to Harrison Bay, 105 miles southeast of Barrow, have ratios of from 3:1 to 4:1. The ratio is considerably smaller in the far western part of the Coastal Plain and approaches 1:1 along the south border and in the plain east of the Itkillik River.

Depth.—The oriented lakes range in maximum depth from about 2 to 20 feet or more. Although the depths of the existing deeper lakes have not been determined, measurements of drained lake basins formerly occupied by oriented lakes show that some were at least 60 feet deep.

As a preliminary and tentative classification based on depth and underwater profile, the lakes may be divided into two groups. One group consists of those lakes which have a shelf or shallow underwater bench surrounding a deeper central portion (pls. 1, *B* and 2, *B*). The other consists of those lakes which are relatively shallow throughout and whose underwater profile is uniformly concave. The two types are usually distinct, although gradational forms exist and absolute classification cannot be adhered to for all lakes. Those few lakes for which underwater data are available and whose maximum depth is less than 6 feet belong clearly in the second group, and those exceeding 6 feet in maximum depth belong in the first. In general, the shallow lakes with uniform concave profiles are most abundant in the northern part of the Coastal Plain, and the deeper lakes with shallow shelves are more abundant in the southern part.

Shore features.—As seen from the air, the major outline of many oriented lakes is smoothly curved, but in detail all shorelines are cuspate, uneven, or jagged.

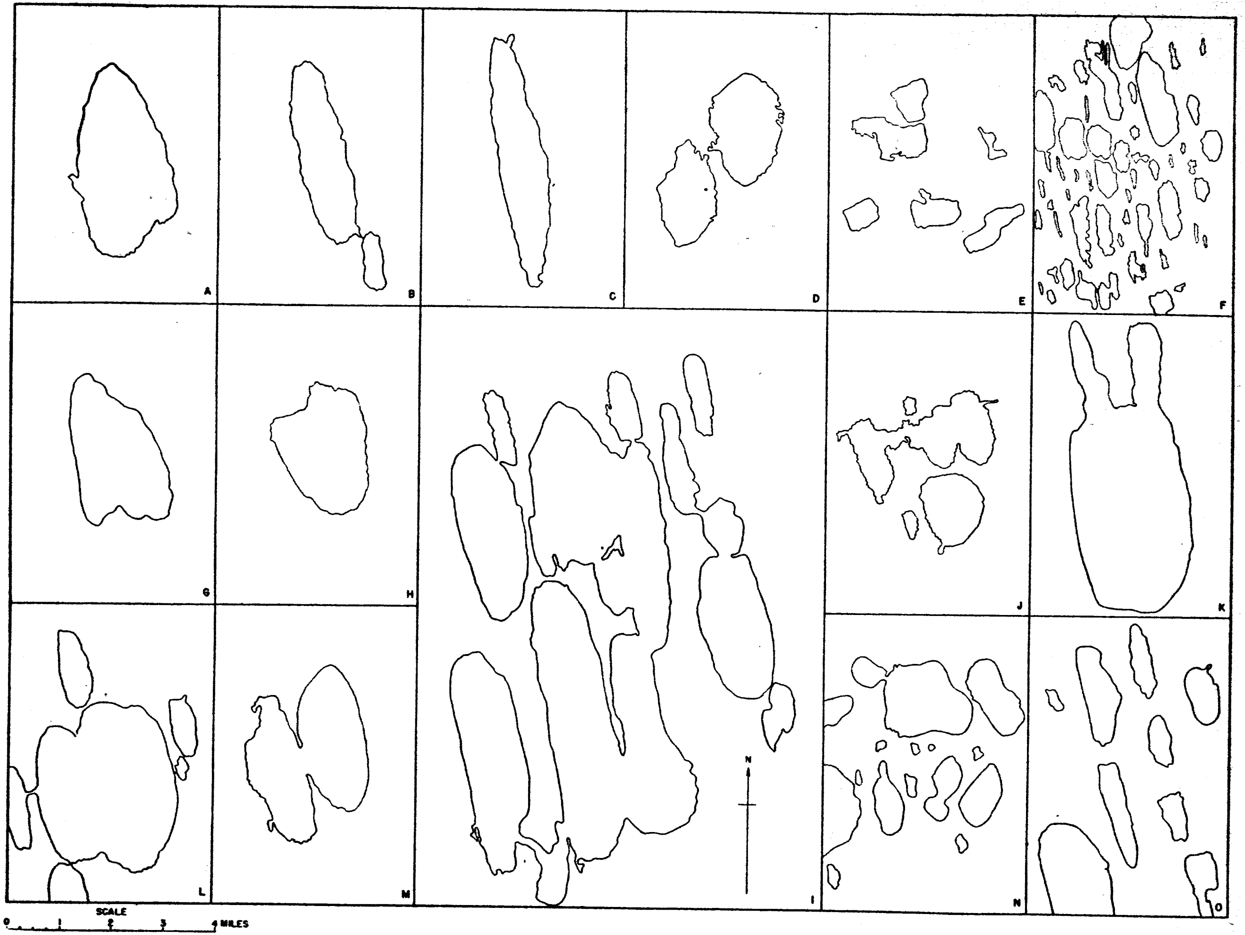


FIG. 4.—Representative outlines of oriented lakes, showing size, shape, and orientation

The degree of unevenness varies from one part of a lake to another and from lake to lake throughout the Coastal Plain. In general, the western shorelines of the oriented lakes are the most uneven, and the eastern shorelines slightly less so. Although the northern and southern shorelines are less smoothly curved in major outline than are the western and eastern, they are generally not so rough or jagged in detail. Relatively smooth, curved shorelines are shown in plate 2, *A*, and cuspate or jagged types in plate 2, *B*.

From the ground, the unevenness can be clearly seen in some lakes (p. 5, *B*), but it is less pronounced in others (pl. 5, *A*, *C*, *D*).

The height of the shore may be different from one part of a lake to another and from lake to lake throughout the Coastal Plain. Generally, the highest shores are on lakes in the southern part of the Coastal Plain. Low wave-cut bluffs range from less than 1 foot to 20 feet above water level and are composed of sand (pl. 5, *D*) or silt, clay, and organic material (pls. 5, *B*, *C*; 6, *A*). In other parts of the same lake the shoreline grades into marshes (pl. 5, *A*). Beaches border some shorelines at low water or when winds are blowing the water away from shore (pl. 6, *A*). As far as is known, no lacustrine beach ridges occur around any lake on the Coastal Plain, but all lakes are incised below the general level of the tundra (pl. 6, *B*, *D*).

FORMER LAKE BASINS

Drained lake basins and former extensions of the present lakes are indicated by wave-cut and wave-built shores, beaches, lacustrine deposits, character of polygonal ground, and character and distribution of vegetation. The evidence shows conclusively that most of the present lakes were much larger than they are today and that many formerly oriented lakes are now completely drained. The abandoned lake basins are clearly visible from the air and in aerial photographs (see pls. 1–3, 4, *A*), and less clearly from the ground. Polygonal ground, commonly in a rectangular pattern emphasized by vegetation, parallels and outlines the old shorelines.

The degree of dissection of the old lake basins varies widely. Some are occupied almost entirely by lakes; others are completely drained and almost obliterated by surface erosion. The former lakes were parallel to one another and commonly overlapped (pl. 3, *A*). The present lake may occupy all, or nearly all, the lake basin; it may be much smaller but similar in shape and orientation; it may be smaller and have no resemblance to the original basin; or it may consist of several ponds, oriented or unoriented, occupying parts of the original basin (fig. 5).

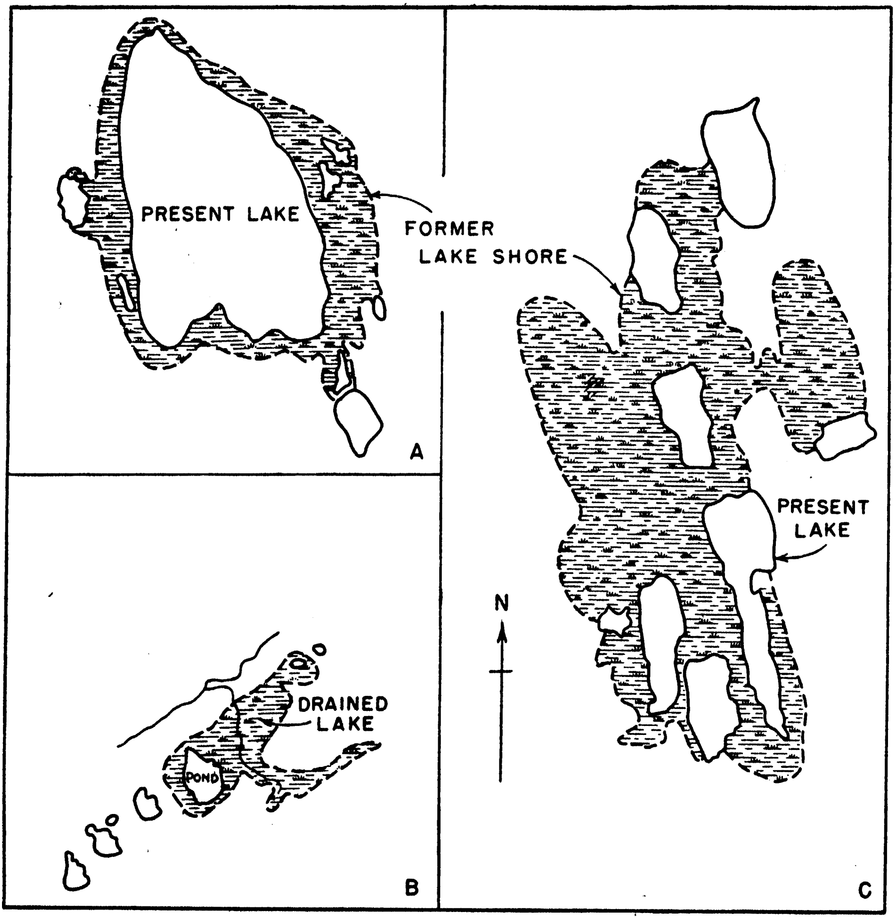
EFFECT OF WIND, VEGETATION, AND PERMAFROST ON THE LAKES

All the oriented lakes are being modified by wind, vegetation, and permafrost.

The dominant east-northeast and west-southwest wind directions are demonstrated by weather records at Barrow, by alignment of dune sands and blowouts along the west and east sides of rivers, and by wave erosion along the west and east sides of the lakes (pl. 3, *B*).

The shallow lakes, those less than about 6 feet in depth, are directly enclosed by permafrost. Under the deeper lakes thawed zones extend possibly to a few yards below lake bottom, but around the shore permafrost is only a few inches deep. In spite of the short cloudy summers, the waters of many lakes warm up to 40° or 50° F., and thawing accompanied by wave action on the frozen sands, silts, and mucks in the banks of the lakes produces very rapid erosion of the shores during storms. Caving, slumping, and erosion of as much as several tens of feet of the shores of some lakes may occur in a few weeks during the summer months. This process results in the cuspate, uneven, or jagged shoreline so common around the lakes and along the coast (pls. 1, *B;* 2, *B;* 3; 5, *B* and *D*) and permits one lake to intersect or capture other lakes (pls. 3, *A*, and 4, *A*).

In contrast to the enlarging effects of thawing of permafrost and subsequent wave erosion, vegetation (commonly *Colpodium fulvum*) in shallow parts of the



REPRESENTATIVE OUTLINES OF EXISTING LAKES AND THE BASINS THEY OCCUPY

SCALE
1 0 1 2 MILES

FIG. 5

lakes tends to reduce and finally to eliminate wave erosion, to fill in parts of the lake (pl. 5, *A*) with plant material, to insulate and protect permafrost from thawing and caving, and to aid in the building-up of permafrost.

These processes go on simultaneously in different parts of a lake or at different times in the same part of a lake. Either destructive or constructive processes may predominate at any one time or period, and apparently they follow cycles connected with such variable factors as temperature, rainfall, cloudiness, wind direction and velocity, and cosmic conditions.

The present tendency is toward destruction of the northerly orientation of the lakes and toward an easterly orientation in those lakes with sufficient "reach" for wave action to be effective. Many lakes are enlarging rapidly as permafrost thaws and waves remove the unconsolidated material forming their shores. These lakes are also becoming shallower, and those too shallow or too small for wave action to be effective are being filled by vegetation. Other lakes are being destroyed by dissection accompanying thawing of permafrost.

In general, the weather data and field observations indicate that the cold reserve of permafrost on the Coastal Plain of northern Alaska is building up by the removal of heat from the ground (Black and Barksdale, 1948; and Black, 1947).

COMPARISON OF THE ORIENTED LAKES OF NORTHERN ALASKA WITH THE CAROLINA BAYS AND OTHER ORIENTED LAKES

As far as is known, this paper contains the first description of a group of oriented lakes which compare in number and in extent of orientation with the Carolina Bays of the eastern United States. Up to the present time the Carolina Bays have been unique, for nowhere else had such a large area containing such well-oriented lakes been known.

The Carolina Bays have been discussed at great length by Douglas Johnson (1942), and it is from his book that data are derived for the comparison with the oriented lakes in table 2.

Obruchev (1940) reported that rectangular lakes oriented in a rectangular pattern occur on the Recent Arctic Coastal Plain to the southwest of the Iamskaia beach of the Okhotsk Sea, in the valley of the Penzhina River near the village of Penzhino (pl. 6, *C*), and in the valley of the Anadyr River. These lakes are reported to vary in diameter from several hundred meters to as much as 5 km. Those on the plain surrounding the Okhotsk Sea are parallel to the coast. Obruchev believes that the origin of the lakes in the valley of the Penzhina and Andyr rivers is connected with the origin of the polygonal ground which occurs in the area, but he does not explain further. He believes that many of the lakes less than 2 meters in depth are the result of the thawing of permafrost. Some lakes deeper than 2 meters are fed by fresh-water springs and others by mineralized waters from *taliks*, or thawed zones within the permafrost. Seasonal and biennial mounds are reported to rise from the bottoms and from the shores of some of the lakes. It is not known how well the lakes are oriented or how large an area they cover. They are presumed to be Quaternary in age.

Johnson (1942, pp. 318–325) briefly discusses oriented lakes of undetermined origin in Florida, Georgia, and Texas and comes to the conclusion that they are similar to, but not identical with, the Carolina Bays.

THEORIES OF ORIGIN OF THE ORIENTED LAKES

The data on the oriented lakes of northern Alaska are too few to permit any definite analogy with the Carolina Bays or with other groups of oriented lakes. However, the Carolina Bays and the oriented lakes of northern Alaska are so strikingly similar that it is believed that the conditions under which they were formed and the physical processes operating to produce them must have been, at least in part, similar. An analogy may be drawn also between the oriented lakes of northern Alaska and the rectangular lakes of eastern U.S.S.R., although data for this comparison are much less satisfactory.

Many conflicting and widely divergent hypotheses are advanced by various

TABLE 2

COMPARISON OF THE ORIENTED LAKES OF NORTHERN ALASKA WITH THE CAROLINA BAYS

	Oriented Lakes of Northern Alaska	Carolina Bays
Distribution	Total area of more than 25,000 sq. mi.; limited to the Arctic Coastal Plain	Total area of at least 25,000 sq. mi.; limited to the Atlantic Coastal Plain
Abundance	Probably tens of thousands of lakes or lake basins	Probably tens of thousands of bays and possibly hundreds of thousands
Orientation	Most trend from N. 9° W. to N. 21° W.; average trend is N. 12° W.	Majority trend from N. 10° W. to N. 55° W.; average trend varies according to locality
Size	A few tens of feet in greatest diameter to 9 miles long and 3 or more miles wide; many average between 1 mile and 3 miles long by ½ mile wide	A few hundred feet in greatest diameter to 4 miles long by 2 or more miles wide
Shape	Elliptical Cigar-shaped Rectangular Ovoid or egg-shaped Irregular Triangular Compound	Elliptical, many with one side more strongly curved than the other Ovoid or egg-shaped Irregular
Ratio of length to width	From almost 1:1 to 5:1; many are 2:1 to 3:1	From almost 1:1 to 3½:1
Depth	Exclusive of filling, from less than 10 feet to 60 or 70 feet below surrounding plain	Exclusive of filling, from 1 or 2 feet to 30 or 40 feet below surrounding plain
Shorelines	Major outline of many lakes is smoothly curved; in detail, most lake shores are jagged or rough.	Shorelines are smoothly curved
Sand rims	Conspicuously absent	Conspicuous, partial, or complete rim or rims of white or pale buff sand around most, but not all, bays
Drainage	No present subsurface drainage; surface drainage absent from many	Subsurface drainage for all; surface drainage absent from many
Relationship to one another	Lakes parallel, intersect, or overlap one another	Bays parallel, intersect, or overlap one another
Relationship to beach ridges	Clusters of lakes lie between what are interpreted as beach ridges	Clusters of bays lie between what are interpreted as beach ridges
Linear arrangement of lakes	Rows of lakes are present in a few localities	Rows of bays are present in a few localities
Material in Coastal Plain	Permanently frozen gravel, sand, silt, or muck	Sands, sandy loams or marls, limestones, and arkosic sand
Present wind direction in relation to axial trend	Average wind is from northeast, normal to orientation of lakes	Winds highly variable and cannot be correlated with trends of bays
Climate at time of origin	Presumably arctic climate similar to that of today	Cold accompanying maximum Wisconsin glaciation
Age	Probably Pleistocene to Recent	Pleistocene

authors to explain the many perplexing problems involved in the origin of the Carolina Bays. Johnson (1942) gives complete references to the numerous papers and discusses all hypotheses. Three of the most outstanding hypotheses are: (1) meteoritic origin, by F. A. Melton, William Schriever, *et al.;* (2) segmented lagoons and crescent-shaped keys, by C. Wythe Cooke; (3) artesian-solution-lacustrine-aeolian hypothesis, by Douglas Johnson.

Of these, the meteoritic hypothesis caused the greatest excitement in the scientific world, but, with certain exceptions, it is not now so widely believed as before. The fact that the oriented lakes are limited only to the Arctic Coastal Plain and almost completely cover that plain, that magnetometer and geophysical surveys do not indicate any deposits of meteorites, and that the shapes of the lakes and absence of "rims" do not conform to such a theory points to the conclusion that the meteoritic hypothesis does not explain the origin of the oriented lakes of northern Alaska.

With some modifications, Obruchev's "cave-in" hypothesis, Cooke's segmented-lagoon hypothesis, and Johnson's complex hypothesis probably will explain some of the problems connected with the origin of the oriented lakes of northern Alaska.

The writers believe that many lakes of northern Alaska, some oriented and others unoriented, are the result of thawing of perennially frozen ground and are true "cave-in" lakes. These lakes may or may not occupy parts of the basins of oriented lakes now drained, and they are possibly of more than one generation. "Caving" in itself, however, cannot produce regional orientation.

In the vicinity of Peard Bay and Harrison Bay and at other localities on the Coastal Plain of northern Alaska, clusters of lakes lie between what have been interpreted as beach ridges (pl. 4, *B*) and are aligned in rows that may have any orientation relative to the axial trend of the lakes. As in the Myrtle Beach area in South Carolina, described by Cooke, such lakes are believed to be the result of uplift and segmentation of lagoons, although the actual processes involved in such segmentation are not clearly understood.

Prevailing winds in the direction of the long axis of the Alaskan lakes are believed to be the chief factor that controlled their orientation. What climatic changes would have to be made to produce a northwesterly (or southeasterly) wind are not known, but presumably they were connected with the major climatic changes during the Pleistocene period and possibly with the most recent uplift of the Brooks Range to the south.

Johnson's (1942) hypothesis of the origin of the Carolina Bays "supposes that artesian springs, rising through moving ground water and operating in part by solution, produced broad shallow basins occupied by lakes, about the margins of which beach ridges were formed by wave action and dune ridges by wind action." Such an origin for the oriented lakes of northern Alaska presupposes that the ground was thawed to a considerable depth during the time that the lakes were forming. Whether during one of the warmer interglacial periods known to have occurred in Alaska (Capps, 1932, and Taber, 1943) the ground was thawed sufficiently to permit circulation of ground waters or artesian waters is not known. Evidence for such phenomena should be looked for particularly in the deeper lakes in the southern part of the Coastal Plain. During such a period of thaw, solution might have taken place in Cretaceous limestone beds, some as much as 6 feet thick and totaling about 100 feet in thickness, which are believed to underlie the area. However, the presence of frozen and well-preserved plant and animal remains dating back into the Pleistocene indicates that there has been no long period of thaw since that time.

Whatever the origin, the lakes present an intriguing problem that deserves considerable attention.

ACKNOWLEDGMENTS.—Many thanks are due Thomas G. Payne and other members of the Alaskan section of the United States Geological Survey who offered suggestions and criticisms based on their knowledge of the geology of northern Alaska. Air transportation from Fairbanks to Barrow and return was furnished by Wien Alaska Airlines under contract with the United States Navy. Arctic contractors provided lodging, meals, and transportation in Naval Petroleum Reserve No. 4.

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THAW LAKES AND THAW SINKS IN THE IMURUK LAKE AREA, SEWARD PENINSULA, ALASKA[1]

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ABSTRACT

Certain lakes and depressions in the Imuruk Lake area, Alaska, are ascribed to subsidence following the thawing of perennially frozen ground. The frozen, silty soils of the region contain large quantities of clear ice, which in volume greatly exceed the natural porosity of the unfrozen material. Melting of the clear ice results in surface subsidence; water accumulates in the resulting depressions.

Thaw lakes are described, and mechanisms of enlargement and eventual drainage are discussed. The origins of drained thaw lakes and of thaw sinks are compared.

Evidence is presented to show that the present climate in the Imuruk Lake area is sufficiently cold to form a small thickness of perennially frozen ground in previously unfrozen deposits but that the present large thickness of frozen ground probably is unstable under existing climatic conditions.

INTRODUCTION

During the summers of 1947 and 1948 thaw lakes and thaw sinks were studied in the Imuruk Lake area, Seward Peninsula, Alaska (fig. 1). The study was made in connection with the permafrost program of the United States Geological Survey and was financed in part by the Engineer Intelligence Division, Office of the Chief of Engineers, United States Army. Mapping was done on aerial photographs made by the Forty-sixth Photo Reconnaissance Squadron, United States Army Air Forces, during 1946.

DEFINITIONS

Thaw depressions.—Depressions which result from subsidence following the thawing of perennially frozen ground.

Thaw lakes.—Lakes which occupy thaw depressions; the term includes lakes which have originated in other ways but which have been considerably enlarged by thawing and caving at their margins (synonymous with "cave-in lakes") (Wallace, 1948, p. 171).

Thaw sinks.—Closed depressions with subterranean drainage; believed to have originated as thaw lakes.

GEOGRAPHIC AND GEOLOGIC SETTING

Imuruk Lake is near the center of the Seward Peninsula, Alaska, at latitude 65°35′ north and longitude 163°10′ west. It is 100 miles northeast of Nome, Alaska, and 60 miles south of the Arctic Circle. The area is a broadly rolling upland, ranging from 500 to 2,000 feet in altitude. Hills in the area are characterized by broad, domical summits and smooth, gentle slopes, rarely steeper than 10°.

An integrated stream pattern is lacking throughout much of the area. Low precipitation, low relief, irregular initial topography, and the disturbing effects of perennially frozen ground and frost action combine to inhibit organization of the drainage. Most of the streams are small and sluggish. Valleys, swales, and low-lying areas generally are swampy. Standing water is common on flat hilltops and in minor irregularities on the gentle slopes. Lakes and ponds are abundant.

Year-round weather observations are lacking in the Imuruk Lake area. The following remarks are based upon weather data collected by the Eleventh Weather Squadron at Lava Lake at the western margin of the area from April through November, 1945, and upon fragmentary weather records kept by the Geological Survey field party during the summers of 1947 and 1948.

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[1] Published by permission of the Director, United States Geological Survey. Manuscript received December 14, 1948.

The climate is rigorous and continental, characterized by cool summers and very cold winters. The mean annual temperature probably is about 25° F. Subfreezing temperatures predominate from early October to mid-May, although there probably are thaws during all the winter months. Nightly frosts are frequent during all the summer months except July. The mean annual precipitation probably is less than 10 inches. The total precipitation recorded from April through November, 1945, was 6.7 inches, of which about 1 inch fell as snow. Eighty per cent of the precipitation was recorded during July, August, and September.

Most of the area is underlain by basaltic flows of Quaternary age. Granite, schist, and schistose limestone are found on the fringes of the area, however, and in isolated hills within the lava plateau. Peat and lacustrine silts from 5 to 20 feet thick are present in terraces on the margins of Imuruk Lake. Stream-laid sediments, chiefly silts, extend to an unknown depth in the floodplain deposits of several of the larger streams. The area has not been glaciated.

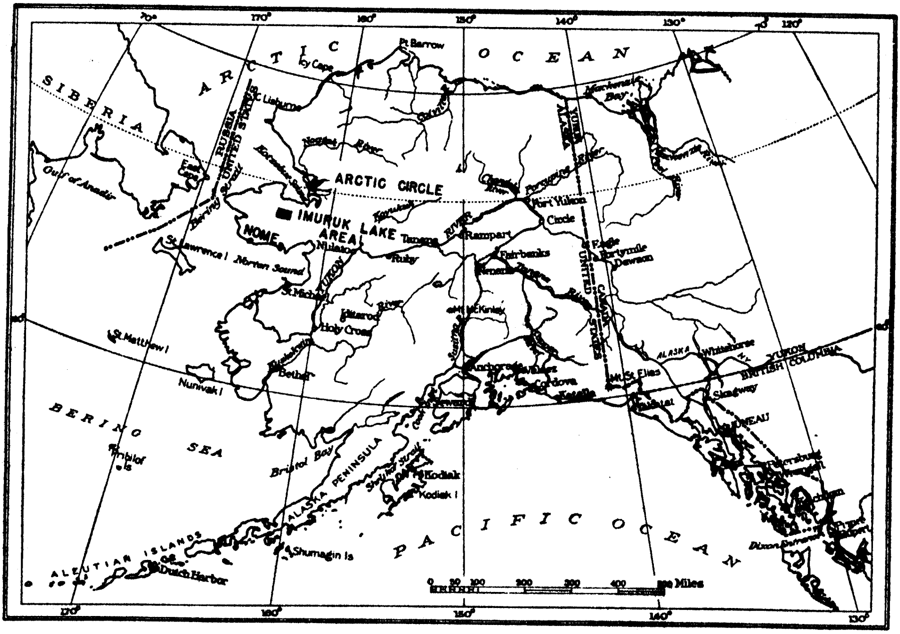


FIG. 1.—Index map of Alaska showing location of Imuruk Lake area

FROST RIVING AND SOIL FORMATION

Previous descriptions of thaw lakes and related thaw features have dealt largely with lakes in areas underlain by a considerable thickness of unconsolidated, fine-grained sediments (Cabot, 1947;

Wallace, 1948). Such sediments have only a limited distribution, however, in the Imuruk Lake area; most of the thaw features occur, instead, in the thick residual soil, which mantles bedrock throughout much of the area. Inasmuch as the characteristics of some of the thaw features depend upon the nature of the soil and of the underlying bedrock, a brief discussion of the soils and the processes that produced them is desirable.

Frost riving is the chief factor in rock decay in the Imuruk Lake area. Two scales of riving can be distinguished: large-scale riving, which takes place along megascopic fractures, such as joints, bedding planes, and foliation planes, and small-scale riving, which proceeds along submegascopic planes of weakness, such as mineral-grain boundaries and cleavage planes within individual mineral grains.

Large-scale riving breaks solid rock into a rubble of fragments ranging from 0.2 to 20 feet in largest dimension. The coarseness of the rubble depends upon the abundance of fractures inherent in the parent rock. Except in certain exceptionally favorable environments, large-scale riving is not an important process in the area today. Widespread stabilized rubble fields indicate, however, that large-scale riving was an extremely active process in the past, probably during the cold stages of the Pleistocene epoch.

Small-scale riving reduces rock fragments to material in the silt and sand grades. The grain size of the resulting debris depends upon the texture and mineral composition of the parent rock. Small-scale riving probably was an important soil-making process in the Imuruk Lake area throughout Pleistocene time and is now the chief active one.

The Quarternary basaltic flows are cut by columnar joints at intervals of 1–6 feet and by more closely spaced, vertical and horizontal fractures which resulted from rapid surface cooling and from continued movement of fluid lava at depth after a surface crust had solidified. Large-scale riving reduces solid basalt to a rubble of 0.5–15-foot blocks. The end-product of small-scale riving in basalt is rather uniform material in the silt range.

The youngest of the Quaternary flows is believed to have been extruded within the last two or three thousand years; the flow has not been subjected to extensive large-scale riving, and its surface consists of solid rock, bearing 1–6 inches of silty or sandy soil in protected hollows and crevices. Slightly older flows have been exposed to the more rigorous climate which accompanied Wisconsin glaciation and have been reduced to extensive rubble fields.

The oldest and most widespread Quaternary flows have been exposed to frost riving throughout much of Pleistocene time and are mantled with 5–20 feet of uniform, fine-grained residual soil. Particles between 0.1 and 0.01 mm. make up 70–95 per cent of these soils at the surface. Scattered boulders, 1–3 feet in diameter, lie on the soil surface and increase in abundance at depth, grading downward into coarse frost-rubble at the base of the weathered zone. A typical cross section showing the gradation from bedrock through frost-rubble to uniform silty soil is shown in figure 3.

DISTRIBUTION AND CHARACTER OF PERENNIALLY FROZEN GROUND

Perennially frozen ground is present throughout the Imuruk Lake area. Drill-hole data and comparisons with other areas having a similar climate indicate that the frozen layer extends to depths ranging from 50 to 300 feet. The widespread occurrence of subsurface drainage suggests, however, that many thawed zones exist within the perennially frozen layer.

Fine-grained sediments or soils, when subjected to subfreezing temperatures in the presence of abundant moisture, take up a quantity of water greatly in excess of the natural porosity of the unfrozen material. Taber (1943, pp. 1517–1518) estimates that the average ice content in perennially frozen silts in central and northwestern Alaska ranges from 50 to 80 per cent, although the porosity of the unfrozen silts would be only 20–30 per cent. The excess ice is segregated in clear ice lenses, wedges, and veinlets distributed through the sediment. Taber (1943, pp. 1522–1527) has outlined some of the processes by which segregations of clear ice are formed in silty material.

Masses of clear ice constitute a major component of areas of peat and frozen, fine-grained soils in the Imuruk Lake region. The clear ice occurs in small, closely spaced horizontal lenses ("ice gneiss," Taber, 1943, p. 1512) and in polygonal networks of vertical ice wedges. The horizontal lenses range from 0.1 to 6.0 inches in thickness and from a few inches to a few feet in length. The vertical wedges generally are from 3 to 6 feet wide and extend from the top of the perennially frozen layer to depths exceeding 10 feet (*D* in fig. 2). The polygonal areas enclosed by the ice wedges range from 35 to 60 feet in diameter. Leffingwell (1909, pp. 205–212) describes similar ice-wedge polygons in the Canning River region, Alaska, in more detail.

During the summer, thawing temperatures penetrate to depths ranging from 1 to 10 feet or more, depending upon the vegetal cover and the nature of the soil. Areas mantled with deep residual soil commonly are covered with a dense mat of peat-forming tundra vegetation, including *Eriophorum vaginatum* (cotton-grass), various shrubs and heaths, and *Sphagnum* mosses. The dense, interlocking mat of living and dead vegetation constitutes an effective insulator, beneath which summer thaw penetrates only 1–3 feet. Beneath areas of bare soil, however, seasonal thaw extends to depths ranging from 3 to 10 feet (*A* in fig. 2). Water constitutes a more effective heat conductor than either peat or mineral soil. Especially deep and rapid thaw takes place beneath and at the margins of small pools of standing water (*B* in fig. 2); perennially frozen ground probably has thawed completely beneath the larger lakes and streams in the Imuruk Lake area.

Ice-rich soils and sediments lose considerable volume upon thawing, owing to the melting-out of bodies of clear ice. Areas of locally deep thaw commonly are marked by subsidence at the ground surface. Small areas of exceptionally deep thaw tend to be self-perpetuating and self-enlarging because the resulting depressions tend to hold small pools of standing water or to channelize slope runoff into small streams beneath which thaw is accelerated.

Some of the most spectacular examples of subsidence due to excessive thaw result from the disruption of the tundra mat by human activities. During 1945, a gentle hill slope had been crossed repeatedly by a caterpillar tractor. When first examined in 1948, the route of the tractor was marked by long, swampy furrows indented 3–12 inches in the tundra. Later in the summer a series of sinkholes, 3–5 feet deep, connected by subterranean watercourses, was discovered beneath the tracks in areas where the tractor had traveled directly upslope (pl. 2, *A*). Still

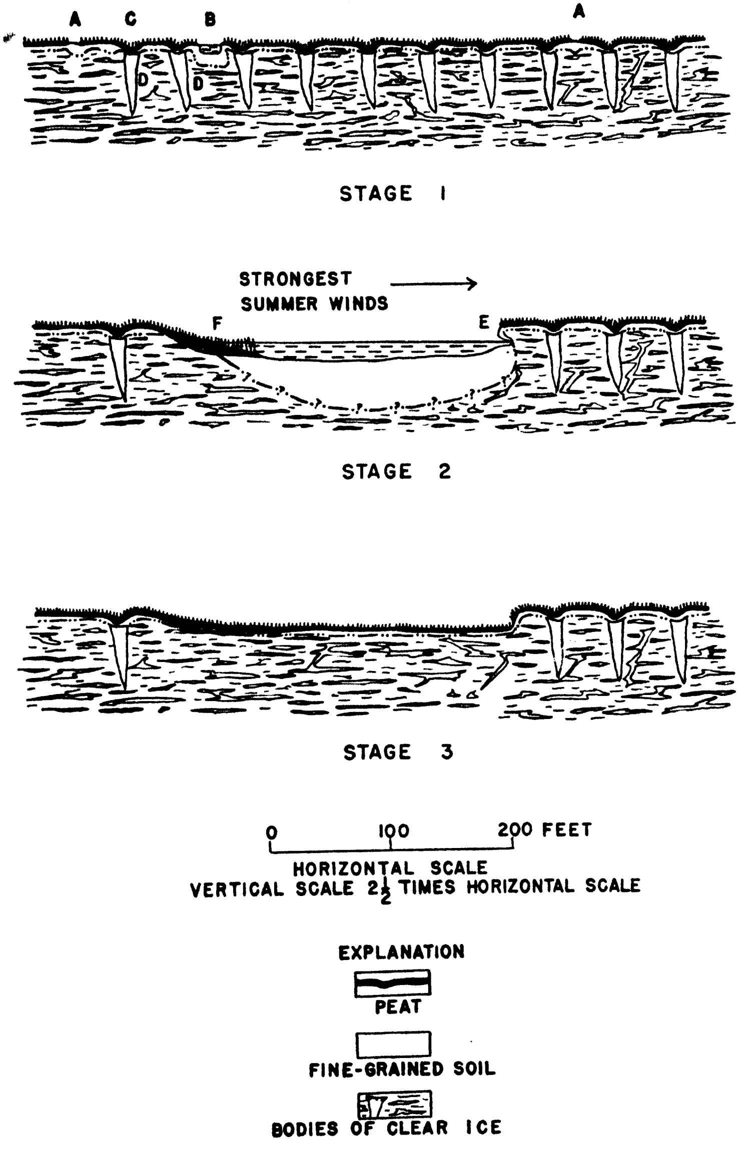


FIG. 2.—Diagrammatic cross sections illustrating the origin and development of a typical thaw lake. Stage *1* shows the initiation of areas and deep thaw beneath frost boils (*A*) and beneath a small pool (*B*). Note shallow trenches (*C*) over buried ice wedges (*D*). In stage *2* the small pool has grown by thawing and caving into a thaw lake, which is migrating in the direction of strongest summer winds. On the windward side (*E*) active caving is in progress, but on the leeward side (*F*) the lake is being filled with peat-forming vegetation. In stage *3* the lake has been drained and new perennially frozen ground has formed beneath the bottom.

later, the roofs of the connecting caverns had collapsed, and the course of the tractor was marked by narrow gullies 3–4 feet deep.

Despite the widespread occurrence of local deep thawing in the Imuruk Lake area, the present climate of the region is cold enough to result in the formation of perennially frozen ground in previously unfrozen deposits. On August 30, 1948, frozen ground was encountered at a depth of 22 inches in a pile of peat 30 inches high which had been plowed up by a bulldozer during 1945. Moreover, new frozen ground has formed in the bottoms of recently drained lakes described below.

THAW LAKES

Lakes ranging from a few tens of feet to several miles in diameter are common in the Imuruk Lake area. Imuruk Lake and many of the larger lakes occupy low areas in initial lava-flow topography. Most of the smaller lakes, however, are thaw lakes, occupying shallow depressions caused by the thawing of perennially frozen ground. These lakes are found on stream floodplains underlain by silty sediments, on stream and lake terraces underlain by peat and silt, and on areas mantled with silty residual soil from 10 to 20 feet thick. The thaw lakes are most abundant on flat surfaces, but they also are present as indentations on slopes as steep as 5°. Most of the lakes are circular or oval in outline; locally, however, several lakes have coalesced to form a single large lake with a more complex, scalloped outline. Those which originated as thaw lakes rarely exceed 1,000 feet in diameter. Enlargement by thawing and caving is a common phenomenon, however, at the margins of larger lakes which have originated in other ways.

The lakes are shallow and flat-bottomed. Water ranges in depth from 1 to 6 feet, and the bottoms lie 10–20 feet below the general level of the surrounding terrain. Many of the lakes are surrounded by a low rampart a few hundred feet wide and about 10 feet high. Such a rampart can be seen on the northern and eastern margin of the drained lake shown in figure 3.

The origin, development, and eventual drainage of a typical thaw lake is shown diagrammatically in figure 2.

In the Imuruk Lake area thaw lakes originate in areas of locally deep thaw which can be initiated in any of the following ways:

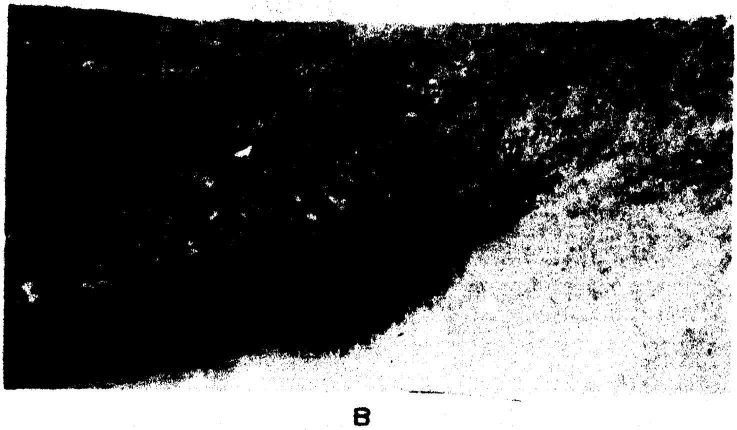
1. By disruption of the vegetal cover by frost heaving. Some of the resulting bare-soil areas subside sufficiently upon thawing to collect small pools of water beneath which thawing is further accelerated (*A* in fig. 3 and *B* in fig. 2, stage *1*).

2. By accelerated thaw beneath pools occupying intersections of ice-wedge polygons. These polygons in the soil are reflected at the surface by shallow swampy trenches (*C* in fig. 2 and *B* in fig. 3). The intersections of the trenches commonly are marked by shallow pools of standing water (*C* in fig. 3).

3. By accelerated thaw beneath pools in small streams. The smaller streams flow in indistinct, vegetation-choked channels. Obstruction of the channels by especially luxuriant vegetation or by small mounds raised by differential frost heaving dams small pools beneath which rapid thawing and subsidence take place. Small streams in areas underlain by deep soil commonly consist of a series of such thaw pools, ranging in depth from 3 to 6 feet, connected by short, shallow-water courses.

Pools of water, once formed, enlarge themselves by thawing and caving at

Partially drained thaw lake



Thaw hole and thaw sink

their margins (pl. 1 and fig. 2, stage 2). Actively caving banks consist of escarpments 3–10 feet high. Thaw proceeds most rapidly at water level; above lake level the banks commonly remain frozen to within 1 or 2 feet of the surface, but, at the water's edge, thaw generally extends several feet into the bank. In places the shores are undercut by cavernous openings at water level, extending 10–15 feet horizontally beneath the banks. Large blocks undercut in this

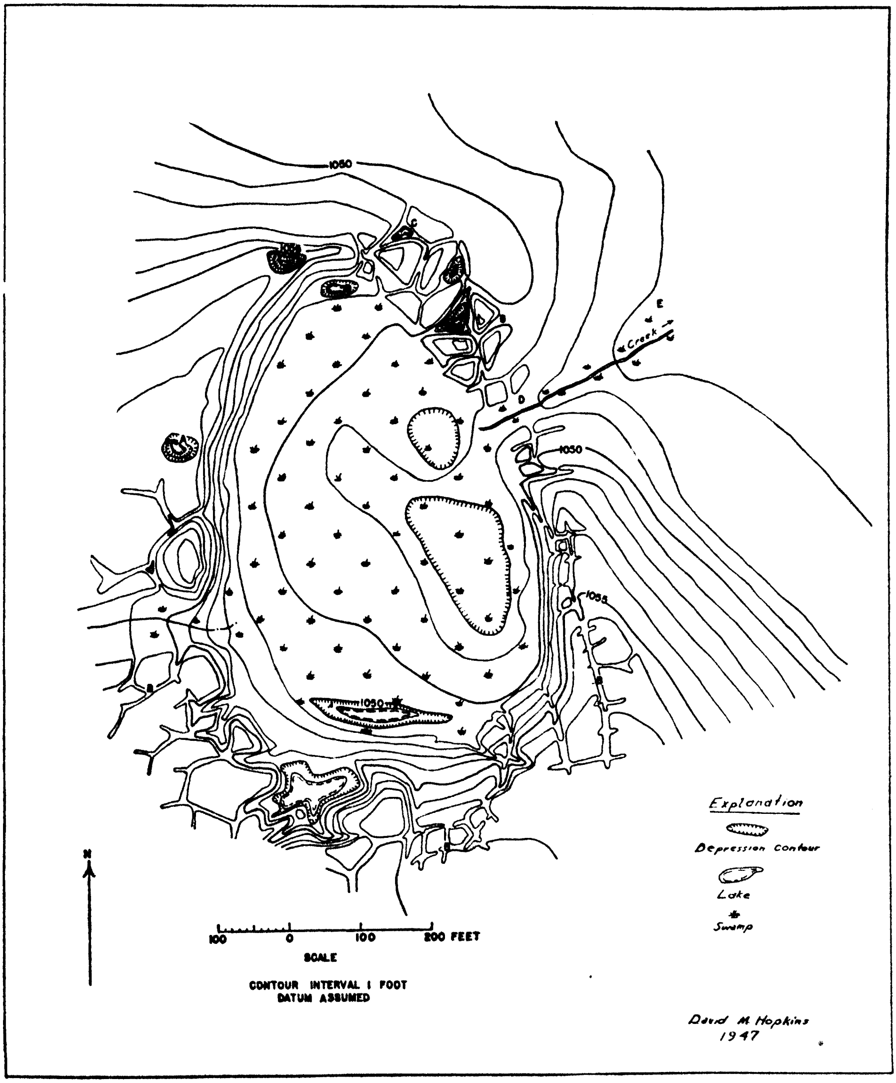


FIG. 3.—Typical drained thaw lake, Imruk Lake area, Alaska. Note small thaw pool at *A*, shallow trenches marking sites of buried ice wedges at *B*, pool at intersection of ice wedges at *C*. The lake probably was emptied following the thawing of an ice wedge at *D*, permitting drainage into the adjoining lowland at *E*.

way occasionally collapse as units. Masses of floating turf, as much as 50 feet in diameter, still covered with living vegetation, were seen near the margins of some of the lakes.

Wave erosion of thawed banks becomes an important process after the lakes have attained a diameter of about 100 feet.

The rate of retreat of actively caving banks varies widely. No perceptible difference was noted between the positions of the shores of many of the lakes, as shown in aerial photographs taken in 1946 and their positions when visited during 1948. At the corner of one lake, however, a retreat of approximately 150 feet had taken place during the intervening 2 years.

Thawing and caving are not limited to the immediate shores of the lakes. Where ice-wedge polygons are present in the banks of the lakes, the bodies of clear ice quickly melt out, and their sites are marked by gullies 2–10 feet deep, extending 50–200 feet into the lake banks (*A* in fig. 4).

If one or more of the ice wedges extend from the lake shore into a near-by area whose surface is lower than lake lev-

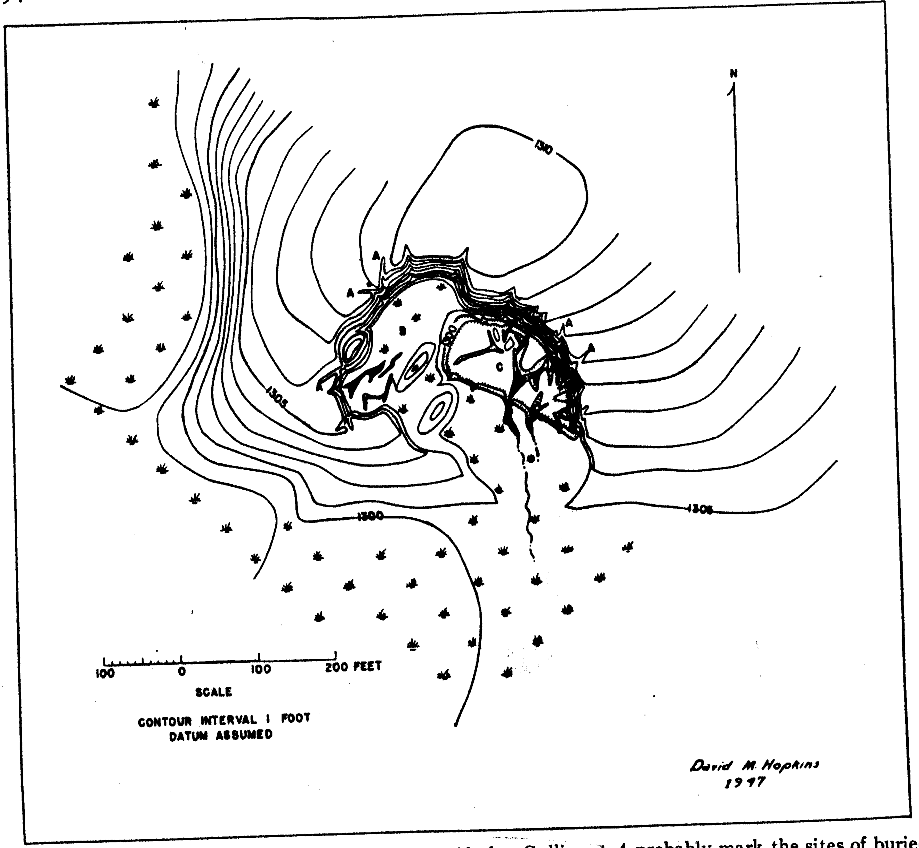


Fig. 4.—Shallow thaw sink, Imuruk Lake area, Alaska. Gullies at *A* probably mark the sites of buried ice wedges. Note flat and marshy area resembling bottom of a typical thaw lake at *B* and hummocky area with polygonal furrows and subterranean drainage at *C*.

el, the trenches formed upon thawing of the ice wedges furnish drainage channels through which the lake eventually may be drained. The level of the lake shown in plate 1 had been lowered several feet in this manner at the time that the photograph was taken. The outlet (*D*) of the drained lake in figure 3 is believed to have originated through the thawing of an ice wedge extending from the lake bottom to the low area at *E*.

Thawing and caving are active around only a part of the circumference of many of the lakes. Where caving is not an active process, the banks are gently sloping or nearly flat. Peat-forming plants, including *Carex* spp., *Eriophorum angustifolium*, and *Sphagnum* mosses, cover such sloping banks and commonly extend 10–100 feet into the adjoining shallow water. Some of the lakes appear to be migrating slowly because of thawing and caving on one shore (*E*, fig. 2) and filling by solifluction and peat accumulation on the opposite shore (*F*, fig. 2). The direction of migration is not the same for all lakes in the Imuruk Lake area. Local topography and the direction of the strongest summer winds appear to be the chief factors in determining the relative positions of caving and filling on any given lake.

DRAINED THAW LAKES

Drained thaw lakes are conspicuous and persistent features of the landscape in the Imuruk Lake area. Newly drained lakes consist of gently undulating areas of bare mud. The bottom is soon colonized by *Carex*, *Sphagnum*, and *Salix*, however, and becomes a flat marsh surrounded by a low escarpment at the site of the former lake banks (fig. 3). Perennially frozen ground forms again in the lake bottom (fig. 2, stage *3*). Evidence of newly formed ice wedges is lacking in the bottoms of drained thaw lakes in the area, although well-developed ice-wedge polygons are present in the floors of drained thaw lakes in the floodplain of the near-by Kuzitrin River. The "rounded polygon areas" on the Arctic Coastal Plain of Alaska illustrated by Cabot (1947, figs. 5 and 6) probably represent drained thaw lakes.

After new perennially frozen ground has formed, the floors of some drained lakes become the sites of smaller new thaw lakes. Several groups of actively growing lakes near Imuruk Lake are surrounded by escarpments marking the former shores of older, larger lakes. Two previous cycles of thawing and caving are recorded 16 miles west of Imuruk Lake, where a group of thaw lakes is surrounded by two sets of such escarpments.

Some of the drained thaw lakes are believed to be very old. A tributary of Noxapaga River has incised itself 25 feet into fluviatile silts in the floor of a large drained lake 13 miles west of Imuruk Lake. The beginning of the present stage of down-cutting on the Noxapaga and its tributaries is tentatively correlated by the author with the beginning of the post-Wisconsin thermal maximum (the "climatic optimum"). If this correlation is correct, the lake must have been drained at least 8,000 years ago.

Certain thaw depressions, closely resembling drained thaw lakes, may never have contained a body of standing water. These features occur chiefly as indentations on slopes of 2°–5°. They are round or oval in outline and range from 100 to 500 feet in diameter. The bottoms generally have a perceptible slope, locally as high as 4°. Slow but active caving is in progress at the margins of the depressions, especially along the courses of ice-wedge polygons. The author was unable to ascertain the cause of the accelerated thaw which produces these depressions.

THAW SINKS

Scattered over areas underlain by the older lavas of the Quaternary basaltic flows is a number of crater-like sinks characterized by interior drainage. The sinks are comparable in size, shape, and distribution to drained thaw lakes but differ in the character of their floors. Typical drained thaw lakes have flat floors with essentially no local relief, whereas the sinks are extremely hummocky, with local relief as high as 10 feet.

The shallower sinks consist of relatively flat-bottomed depressions lined by step escarpments, which probably represent the former banks of thaw lakes. Locally the floors may be flat and marshy, identical with the floors of typical drained thaw lakes (area *B* in fig. 4). In each of the sinks, however, at least a part of the floor consists of a system of polygonal furrows ranging from 3 to 10 feet in depth (area *C* in fig. 4). The pattern and the horizontal dimensions of the polygonal furrows are similar to those of ice-wedge polygons (compare furrows at *C* in fig. 4 with ice-wedge trenches at *B* in fig. 3). Shallow gullies in the walls of some of the sinks (*A* in fig. 4) appear to represent extensions of the furrows on the floors.

Silty basalt soil containing few or no boulders is found in the walls of the sinks. The walls of the furrows, however, are composed of coarse basalt rubble in a matrix of silt. Clean rubble with open interstices is found locally in the bottoms of the furrows. The coarse material in the furrows is believed to represent frost rubble at the base of the soil zone. Figure 5, stage *2*, presents a typical cross section through a shallow sink.

The deeper sinks depart more radically from the form of typical drained thaw lakes. The walls of these sinks slope inward rather uniformly toward a single steep-walled linear furrow, which con-

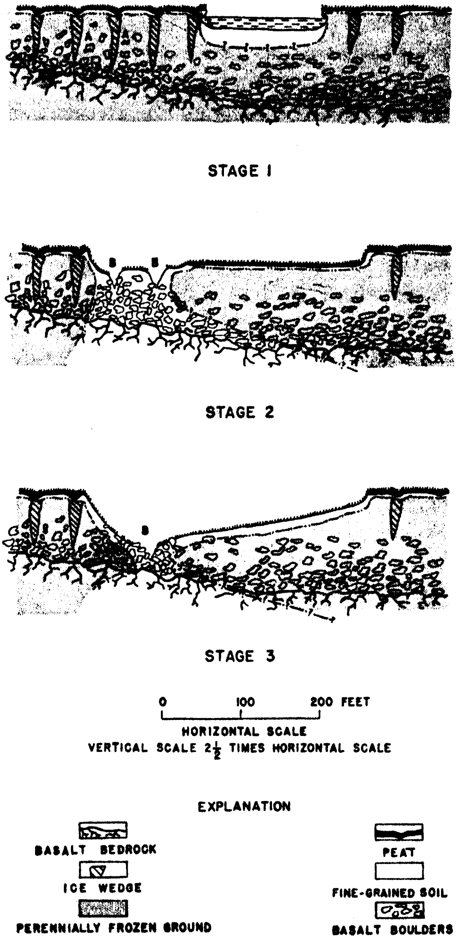


FIG. 5.—Diagrammatic cross sections illustrating the origin and development of a typical thaw sink. Stage *1* shows a thaw lake spreading toward an area in which ice wedges extend to the base of the soil zone (*A*). In stage *2* the deep ice wedges have melted out, and the lake has been drained through frost-rubble at the base of the resulting furrows (*B*). New perennially frozen ground has formed beneath part of the lake bottom. In stage *3* one of the furrows has become the dominant outlet. Considerable fine-grained material has been removed by streams from the lake bottom and redeposited in subterranean channels through which the sink is drained.

stitutes the deepest portion of the sink (pl. 2, *B;* and *AA′*, fig. 6). A diagrammatic cross section through a deep sink is given in figure 5, stage *3*. Shorter, diagonal furrows branch from the main furrow at intervals of 50–150 feet.

All the thaw sinks are drained through the open rubble in the furrows. The watershed of some of the sinks is limited by the walls of the sinks themselves, but other sinks receive the drainage of one or more streams having discharges of several second-feet. In all cases the stream eventually disappears into open rubble in the base of a furrow.

Many of the sinks are occupied by ephemeral lakes. The sink illustrated in figure 6 was visited half-a-dozen times during 1947. During most visits it was dry. At 9:00 A.M. on September 2, 1947, however, the central furrow was filled with water to a depth of 10 feet (pl. 2, *B*). A line of grass and wet mud several feet above the water's edge indicated that the water level had been even higher earlier in the day. At 1:00 P.M. on the same day the pool had drained completely, and the furrow was once more dry. Comparable bodies of standing water were not seen in any of the other sinks visited; but, in many of them, areas of bare soil, mud coatings on boulders, or horizontal lines

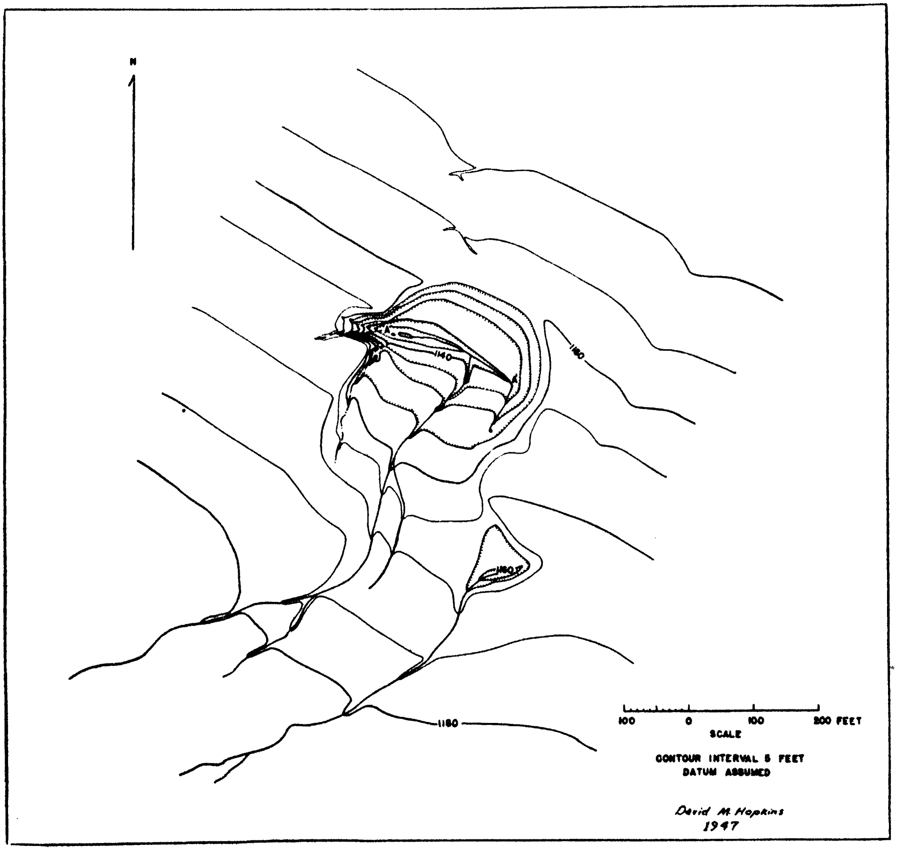


FIG. 6.—Deep thaw sink, Imuruk Lake area, Alaska. Note subterranean drainage into deep furrow at *A–A′*. Same sink is shown in plate 1.

of vegetal debris indicated the former presence of ephemeral lakes.

The similarity in size, shape, and occurrence of thaw lakes and shallow thaw sinks suggests that the sinks originate as thaw lakes. The furrows on the floors of the sinks are believed to be the sites of thawed ice wedges which formerly extended from the top of the perennially frozen layer through the soil zone and into the underlying zone of frost-rubble (ice wedges at *A* in fig. 5, stage *1*). The drainage of a thaw lake to form a thaw sink probably occurs when the lake extends itself over an area of such throughgoing ice wedges. The clear ice masses quickly melt out beneath the lakes, leaving a void extending through the entire thickness of the soil zone (*B* in fig. 5, stage *2*); the water in the lake drains through the void into the underlying frost breccia, thence downward into solid but permeable basalt.

The great depth and the funnel-like shape of the thaw sink (fig. 5, stage *3*) represent modifications of the original thaw lake which could be achieved only by the removal of considerable material from the floor. It is estimated that at least 8,500 cubic yards of soil were removed from an originally flat-floored thaw lake during the development of the thaw sink illustrated in figure 4. The missing material in this and other sinks must have been removed by the small streams entering the ground at the bottoms of the sinks, and redeposited somewhere in the subterranean channels through which the sinks still continue to be drained.

The ephemeral lakes probably originate in several ways. In some sinks, water probably accumulates when mud slumps from the walls of the furrows and covers the open spaces through which the sinks are drained. Sedimentation in the drainage channels by currents overloaded with silt is a possible cause of stoppage of water flow in other sinks. In still other sinks the drainage channels probably fill with ice during periods of diminishing flow in the late fall; a lake then accumulates in the spring until the channels are again thawed.

CONCLUSIONS

The thaw lakes of the Imuruk Lake area originate because of differences in the thermal conductivity of natural materials, including peat, mineral soil, and water, in a region of perennially frozen ground. Inasmuch as perennially frozen ground is found in the floors of drained thaw lakes and thaw sinks, one is not justified in assuming that the presence of thaw lakes and sinks in any particular area indicates that the frozen ground of the area is an unstable relic of a former, colder climate. The presence of the lakes simply reflects the occurrence of a thawing season sufficiently long or warm to produce important differences in the depth of thaw in and beneath bodies of peat, mineral soil, or water.

The widespread occurrence of subterranean drainage in the Imuruk Lake area indicates, however, that the present great thickness of perennially frozen ground is unstable and that the present climate is capable of producing only a much smaller thickness of frozen ground than now exists. Subterranean drainage is possible only if ice is absent from the open spaces which act as drainage channels. Small streams entering the ground during the summer probably are capable of maintaining open channels in rocks in which the temperature remains below freezing throughout the season; with falling discharge in the fall, however, channels in subfreezing zones can be expected to be clogged with ice; when flow ceases, the channel should be completely filled. The occurrence and persistence of widespread subterranean drainage in the Imuruk Lake area indicates the existence of widespread thawed zones within the deeper parts of the perennially frozen layer, which, in turn, suggests that the deepest part of the frozen zone is unstable and gradually thawing in the present climate.

The recognition of thaw features in the field and on aerial photographs is useful in engineering and in water-supply studies. Considerable subsidence can be expected in areas in which thaw lakes, sinks, or depressions are present if the surface vegetal mat is disturbed or if the ground is heated beneath poorly insulated buildings. Thaw sinks indicate the presence of thawed zones in a permeable substratum in which some supply of ground water can be expected.

ACKNOWLEDGMENTS.—Many of the ideas presented in this paper are products of discussions with R. F. Black and Dr. Kirk Bryan and with the author's assistants, E. H. Muller, A. T. Fernald, R. S. Sigafoos, and J. S. Seitz. Grateful acknowledgment is made to Dr. R. R. Shrock and E. C. Robertson for their careful criticism of the manuscript.

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A POSSIBLE FOSSIL ICE WEDGE IN BUREAU COUNTY, ILLINOIS[1]

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ABSTRACT

A wedge-shaped projection of early Wisconsin till into underlying Farmdale loess indicates that the ice advanced over periglacial frozen ground and that a crack, possibly formed by melting of a ground-ice wedge, became filled with till.

INTRODUCTION

Ice wedges are well-known features in present areas of perennially frozen ground (Leffingwell, 1915, pp. 635–654; 1919, pp. 205–212; Taber, 1943, pp. 1510–1516), and casts of ice wedges now filled with sediment are described from permafrost areas in Alaska (Taber, 1943, pp. 1493–1495, pls. 15 and 16) and Baffin Land (Patterson, 1940, pp. 99–102). Similar structures in Pleistocene deposits have been interpreted as fossil ice wedges formed under former periglacial conditions. They are described from numerous localities in Europe (Soergel, 1936, pp. 223–247; Patterson, 1940, pp. 102–107; Weinberger, 1944, pp. 539–544; see the latter paper for additional references) and from one area in New England (Denny, p. 338, figs. 6 and 7). Leffingwell (1919, pp. 205–212, 223–224) attributed the growth of ice wedges to enlargement of original shrinkage cracks by slow growth of ground-ice, and most later writers have adopted this view. Taber (1943, pp. 1519–1527), however, argued against this interpretation and proposed that they were formed simply as segregations of ground water during downward freezing.

The structure described in the present paper is of interest because of the seeming rarity of these features in the United States and because it appears to be the first structure described in which till, rather than loess or water-laid deposits, constitutes the wedge filling.

The till wedge is exposed in a stream cut along a small south tributary of Brush Creek in southeastern Bureau County, about 2 miles northwest of Depue, Illinois (NW. $\frac{1}{4}$, SW. $\frac{1}{4}$, SE. $\frac{1}{4}$, Sec. 28, T. 16 N., R. 10 E.; Hennepin quadrangle map). The exposure lies well within the area covered by Wisconsin drift near the inner margin of the Cropsey moraine (fig. 1). Ice movement during Wisconsin glaciation was westward, more or less normal to the exposed section. Streams in the area have eroded through the Wisconsin till sheet and exposed underlying Iowan loess, Farmdale loess and silt, Sangamon soil, and Illinoian drift.

DESCRIPTION

The Pleistocene section in which the till wedge is exposed is described in table 1. The upper till is identified as Bloomington-Normal drift on the basis of its maroon color, which is characteristic of these deposits throughout western and central Illinois. Cropsey drift, the surficial deposit of the area, does not appear in the section but occurs higher up the slope.

The toothlike wedge of Wisconsin till

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[1] Published with the permission of the Chief, Illinois State Geological Survey. Manuscript received October 20, 1948.

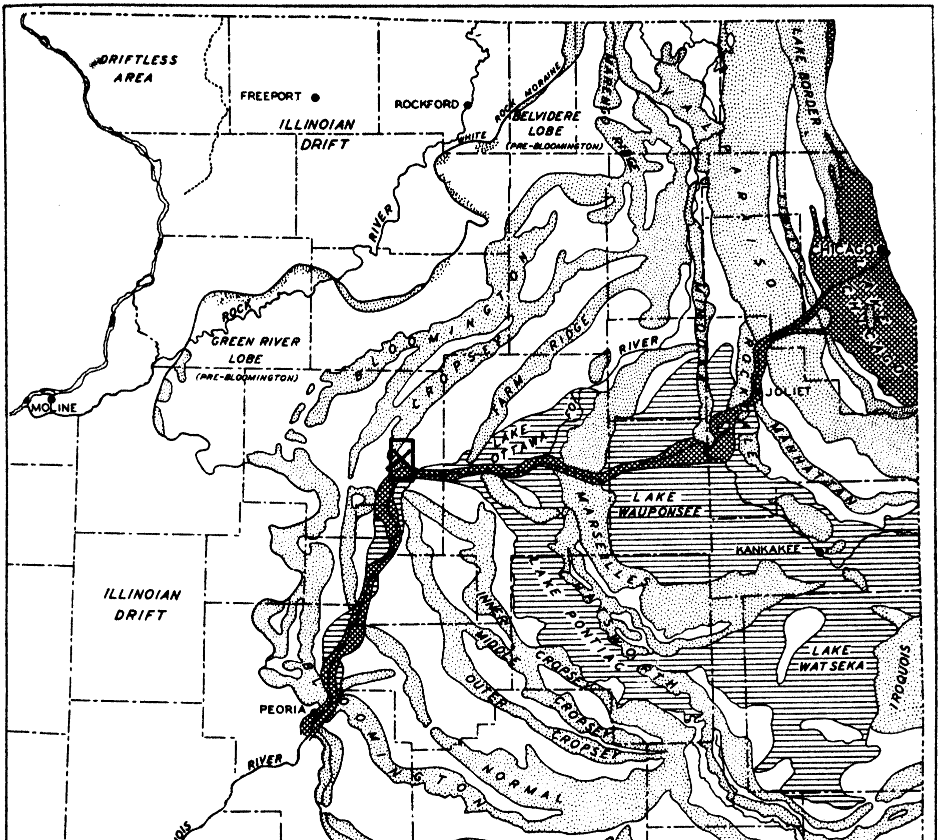

FIG. 1.—Glacial map of northeastern Illinois, showing location (x) of till wedge exposure. Map compiled by George E. Ekblaw, Illinois State Geological Survey, 1942.

projects downward into the Farmdale loess on an essentially vertical face and is about 2½ feet long and 8–12 inches wide (fig. 2 and pl. 1).[a] The outlines of the wedge are angular and clearly are determined by joints in the enclosing loess. The till is uniformly massive and has a blocky appearance due to jointing. In places thin seams of till intrude fractures in the loess. The loess is dense, but an intricate fracture system is revealed by iron-stained joints and inconspicuous breccias composed of loess fragments, which differ slightly in color. Gastropod shells in the loess are broken and fragmented.

TABLE 1

Pleistocene Section on Brush Creek

	Thickness (Feet)
Recent soil and slope wash	
Silt, with gravelly layers at base, yellow	3
Pleistocene series	
Wisconsin drift (Bloomington-Normal)	
Till, calcareous, yellow-gray	6
Sand and gravel, silty, calcareous, yellow-gray	3
Till, calcareous, maroon, dense	4
Farmdale loess (early Wisconsin or late Sangamon)	
Silt, noncalcareous, greenish-gray; weathers maroon-gray to brown; humus in irregular discontinuous layers and flakes; broken gastropod shells; hard, dense, jointed; sharp contact at top, grades into silttil below; penetrated by wedge of overlying till	5
Illinoian drift	
Silttil, noncalcareous, reddish-brown; silicious residuals and ghosts of pebbles, friable; Horizon 2 of Sangamon soil	4½
Till, noncalcareous, yellowish-gray, mottled; Horizon 3	3
Till, calcareous, yellowish-gray; Horizon 4	4
Till, calcareous, dark lead-gray	1

It is probable that the upper part of the wedge was removed by glacial erosion and that the structure originally penetrated a thin layer of Iowan loess overlying the Farmdale loess. The thickness removed, however, was probably less than 5 feet, as the thickness of the Farmdale loess in the section is essentially normal for the area.

INTERPRETATION

The relations indicate that a fracture in frozen loess formed in the periglacial zone of the Bloomington-Normal glacier and became filled with till either during or shortly after glaciation. It is unlikely that an open fracture could be overridden by ice and filled with till except in solidly frozen ground; and, if the fracture were ice-filled, frozen ground would be implied. This condition is indicated also by the widespread preservation of thin Iowan and Farmdale loess deposits below the Wisconsin drift throughout the periphery of this drift sheet in Illinois (Horberg, 1947). These deposits have an average combined thickness of less than 10 feet, and it is improbable that they could have survived glaciation except as frozen ground.

Four possible modes of origin for the original opening may be suggested: (1) ice shove, (2) glacial plucking, (3) shrinkage due to contraction of frozen ground with lowered temperatures, and (4) melting of a ground-ice wedge. The fracturing in the loess suggests ice shove, but such characteristic features as thrust blocks and inclusions of underlying materials in the basal till are absent. Other than the

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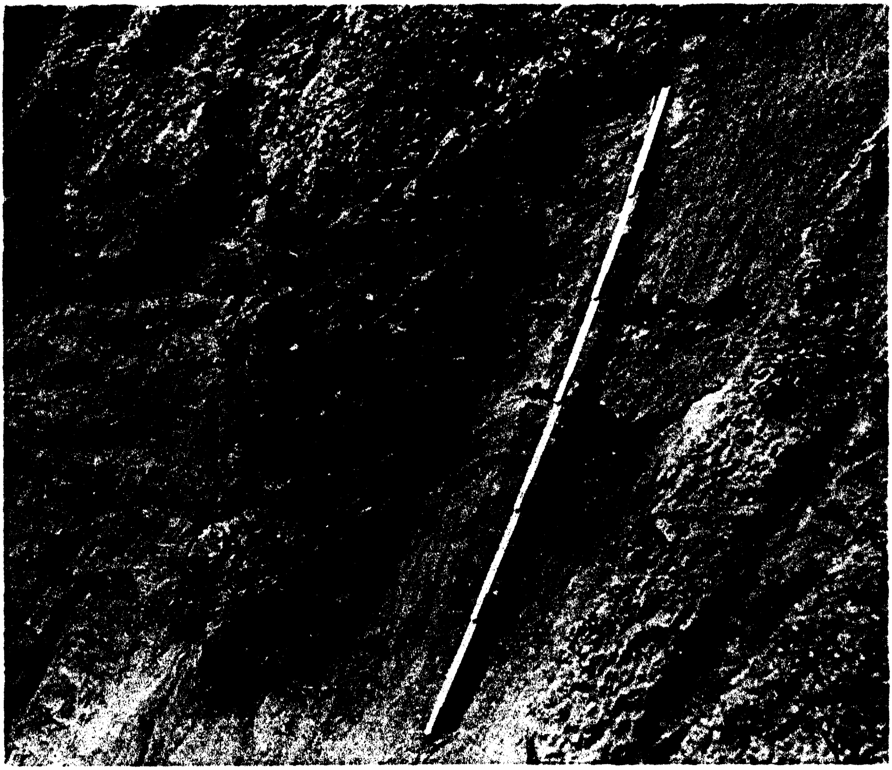
[a] The description is based on observations made during August, 1946. In September, 1948, the exposure face had receded, so that the wedge was reduced to a length of 1½ feet, with some change in its outlines.



Till wedge in Farmingdale loess

wedge itself, there is no indication of glacial plucking. Fractures opened by shrinkage alone, without enlargement by growth of ice wedges, appear to be too small to meet requirements (Taber, 1943, p. 1521). The first three suggestions all raise the problem of how an open fracture could survive glaciation without disruption and become filled with till. To postulate an opening formed by the melting of an ice wedge is much less objectionable, in that the fracture would be filled with ice during glaciation and the till cast would not be formed until after glacial retreat, as the ice wedge slowly melted and overlying ground moraine intruded the opening.

The till wedge differs from many fossil ice wedges in the absence of marginal segregations in the fill and the lack of folding in the enclosing sediments. This may be explained in part by the unbedded character of both the till and the loess. Also, it appears that folding is not conspicuous in the stratified sediments enclosing some of the Alaskan wedges (Taber, 1943, pls. 16 and 17). The size and shape of the till wedge compares closely with the structures described from Alaska and Europe.

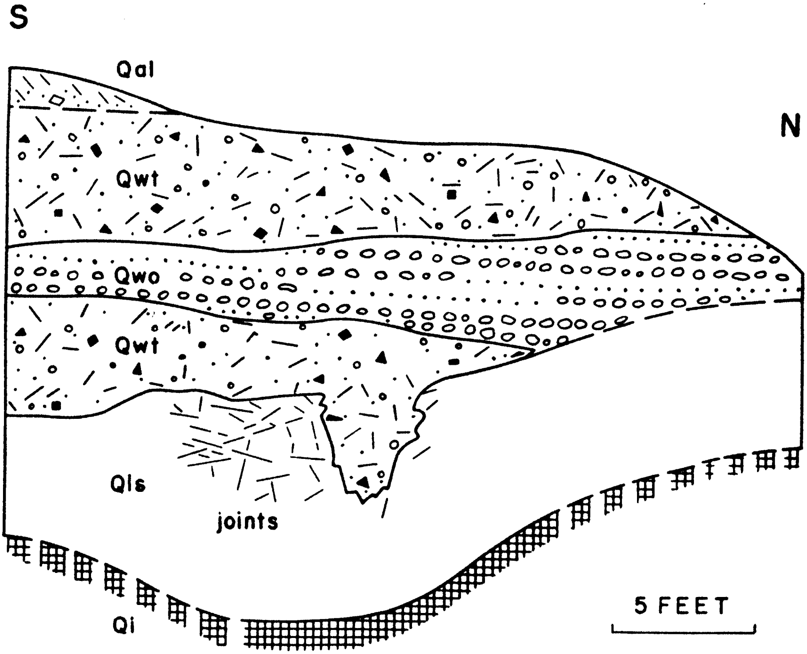


FIG. 2.—Sketch of till wedge, showing stratigraphic relations. *Qi*, Illinoian silttil; *Qls*, Farmdale loess; *Qwt*, Wisconsin till; *Qwo*, Wisconsin outwash; *Qal*, slope wash.

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ROCK-FRAGMENT SLOPES CAUSED BY PAST FROST ACTION IN THE JURA MOUNTAINS (AIN), FRANCE[1]

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ABSTRACT

Partially cemented slopes composed of coarse angular limestone fragments and exhibiting a loose "open-work" texture are described from a portion of the valley of the Ain River in the southern Jura Mountains of France. These slopes are no longer forming but are undergoing erosion by slope wash, creep, and minor sliding. Intense frost action during one or more substages of the last or Würm glaciation is considered to have been the dominant process in their development.

GENERAL STATEMENT

The accumulation of rock-fragments on slopes and in detrital cones is a common and striking feature of the landscape in the southern Jura Mountains of eastern France.[2] These slopes, no longer forming under the present climatic regime of the region, were formed under conditions which prevailed in the immediate geologic past.

Rock-fragment accumulations have been observed over a wide area, including the southern Jura and western Switzerland. Detailed inspection has been confined, however, to that part of the Department of Ain within the Jura Mountains and more particularly to the drainage of the Ain River before it leaves the mountains and joins the Rhone River in the plains of Lyon. As argued below, these slopes are essentially a product of intensive frost action of the past. Other phenomena, including gentle solifluction slopes and landslides, also record this past climate. They are less spectacular, however, than the rock-fragment slopes and are not here considered.

TERMINOLOGY

The material comprising the slopes is locally termed *éboulis*, a French noun meaning "debris," or "fallen rocks." The slopes are often referred to as "*éboulis* slopes," "detrital cones," or "cones of dejection." Cailleux (1948) uses *éboulis ordonnés* for similar deposits. The major components of these slopes are angular fragments, broken from outcrops of Jurassic limestone by frost action and moved downslope. As such they may be grouped under the general term "congelifract," a word proposed by Bryan (1946) to include angular rock fragments produced by frost action. "Congelifract" is used in the following paragraphs to designate the individual fragments within the slope, and "congelifractate slope" is applied to the topographic form resulting from the accumulation of masses of rock fragments or congelifracts.

DESCRIPTION OF SLOPES

EXTERNAL FEATURES

The congelifractate slopes are distributed along the walls of major valleys, along side streams, gulches, and ravines, and around the flanks of many hills.

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[1] Manuscript received October 14, 1948.

[2] The observations here recorded were made during the summer of 1948, while the writer, in the company of Professor Kirk Bryan, Harvard University, was investigating the geologic antiquity of a late Paleolithic archeological site (Abri-sous-Roche de la Colombière) near Poncin, Ain, France, on the Ain River.

Even the steeper portions of the slopes support a natural vegetation or are put into cultivation. All must have been cut over one or more times in the past, and today many are covered with bushes or second-growth trees. This growth may cover the slopes in whole or, if they are used for pasture, at least in part. Vineyards have been planted on many slopes, and a grove of walnut trees is found on one of the steeper slopes near the site of La Colombière. In several cases the slopes are almost completely bare, with only a few scattered bushes and a sparse, discontinuous cover of grass.

Without exception, the deposits display an intimate relation to the bedrock, which crops out upslope and in the past provided the masses of congelifracts which form the slopes. The simple form tends to be fan-shaped and radiates downslope from an apex which butts against a bedrock outcrop. Commonly, a series of such fans may coalesce along a valley wall or hillside. In other places the debris slopes may extend along the foot of a bare rock cliff as a great sheet of rubble with no distinct apex. Most debris cones exhibit a minor bench or shelf at their upward limits along the adjacent bedrock. At some localities the congelifractate slopes are compound, as illustrated in figure 1. Remnants of an old slope, now partially eroded, stand out in front of a more recent and only slightly dissected slope.

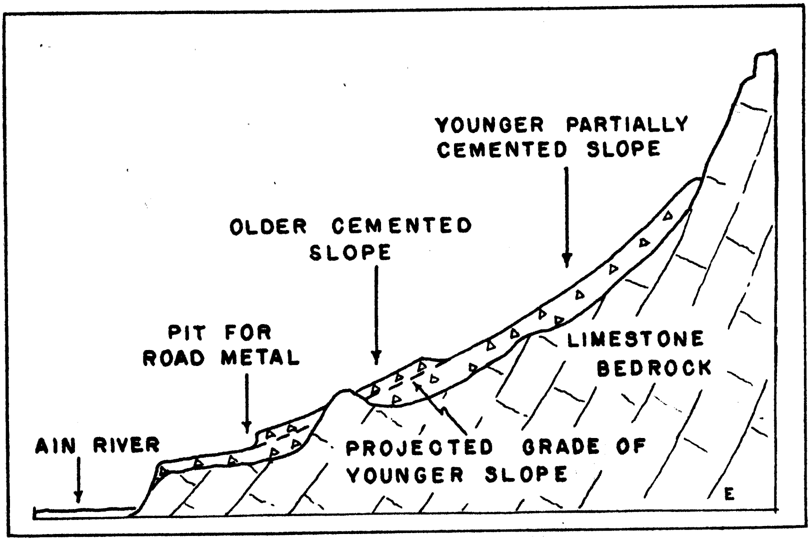

FIG. 1.—Schematic cross section of multiple rock-fragment slopes in the Ain Valley, 1 km. south of Merpuis.

Steepness of slope is characteristic. Figure 2 illustrates profiles of five slopes surveyed by plane table at the rock shelter of La Colombière. They are plotted to a common base, with a vertical exaggeration of twice the horizontal scale. The slopes are concave to the sky and range in mean grades between 17 ($9\frac{1}{2}°$) and 40 per cent (22°). Although no other quantitative data are available on the grade of congelifractate slopes in this area, it seems probable that they average between 30 and 35 per cent (16½° and 19°). Many give the impression of having much steeper grades, an impression undoubtedly due to the fact that one is immediately struck by the steep upper portions of the feature, while the gentler grades downslope go unnoticed.

INTERNAL FEATURES

The *éboulis*, the rock fragments composing the congelifractate slopes, are characteristically angular pieces derived from the bedrock outcrops immediately upslope. They vary in size from approximately 1 inch to 5 or 6 inches in maximum diameter and display a rude sorting. The resultant bedding tends to approximate the angle of slope of the modern detrital cone, although erosion subsequent to deposition has, in many places, cut the bedding at a low angle. Large boulders, up to 2 or 3 feet in diameter and derived from outcrops above the slopes are scattered through the masses of congelifracts. At some localities a few rounded river pebbles are present in the rubble. Their presence is attributed to the destruction of high-level terraces during the formation of the congelifractate slopes and the subsequent inclusion in the rubble beds of some of the component parts of the terrace deposits.

One of the most common and striking features of the internal structure of the congelifractate slope is the relatively low content of the finer-grade sizes. The production of such a great mass of rubble as is represented in the congelifractate slope must have resulted in a fairly high percentage of finer fragments. They are not now present, and it seems reasonable that they were removed penecontemporaneously with the deposition of the congelifractate beds. Abundant supplies of surface water during the periglacial conditions which surrounded the formation of the *éboulis* beds would serve as a mechanism for the removal of the finer particles. Whatever the reason for the removal of fine fragments, their absence results in voids between adjacent individual congelifracts. This lack of interstitial filling produces a characteristic "openwork" texture in the slope deposit, as shown in the photograph in plate 1.

Some fine material does exist in the form of silt- and clay-sized particles and is thought to be the reflection of soil processes operating after the deposition of the congelifracts. These same soil processes impart to any given exposure a general yellow to yellow-brown color. However, a discontinuous zone of white, powdery lime carbonate, 1–2 feet thick, is present somewhat below the modern surface. This zone is followed by additional yellow to yellow-brown staining, which penetrates through the carbonate zone. A few inches of gray to gray-black modern soil form the upper limit of the deposit.

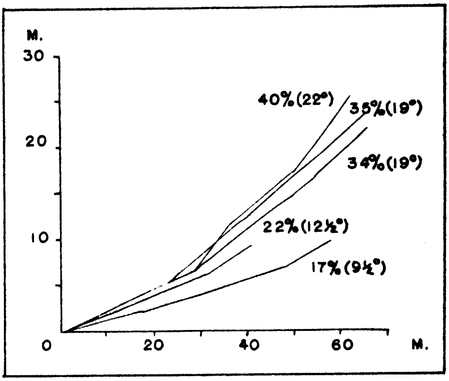

FIG. 2.—Profiles of rock-fragment slopes surveyed at the site of La Colombière, near Poncin. Mean grades in percentages and degrees. Vertical exaggeration 2×.

Varying amounts of cementation by calcium carbonate are found within the slopes. In some places cementation is so complete that the interstices between rock fragments are completely filled and a breccia is formed. More commonly the slope deposits are cemented only enough to permit the *éboulis* to stand at a high angle in artificial cuts and to prohibit extensive and rapid gullying in the slopes.

PRESENT-DAY ACTION ON ROCK-FRAGMENT SLOPES

Intervention by man has modified the congelifractate slope in minor details. Gravel pits for road metal and fill, as well as road cuts, provide many excellent exposures but also contribute to instability of the slope. These artificial cuts, along with the partial destruction of the vegetative cover by grazing, cultivation, and deforestation, have exposed the slopes to present-day erosion. The processes active today consist of essentially minor sliding, creep, and slope wash. Little new material is being added to the slopes, and modern movement is confined to the downward translation of previously deposited congelifracts.

Movement is restricted to a thin surface zone, 1–3 inches thick. Here the calcium carbonate cement has been dissolved by recent soil processes or by rain wash, or the congelifracts have been loosened by the hoe and foot of man or the trampling of grazing goats and sheep. The activity of man and animal promotes the downward sliding of the congelifracts so loosened. Furthermore, rain wash also operates on bare slopes. Where they have been abandoned, heavy rains form broad, shallow runnels. The smaller fragments tend to move downslope, leaving behind the coarser congelifracts in the channels.

The present movement cannot be considered analogous to the processes which produced and deposited the rubble forming the rock-fragment slopes. At no place are rock fragments comparable to those within the slopes forming today. It is true that large blocks still fall from near-vertical cliffs. The small rock fragments, however, have a different aspect from those of the slope deposits. They are essentially thin plates or flakes which scale off the bedrock outcrops and are not the sharp, angular fragments of the congelifractate slopes. That the slopes are not forming today is also evidenced by the narrow shelf or bench found at the upward limit of many slopes next to the bedrock outcrop. This must be a secondary feature formed by erosion after the formation of the debris cones and under conditions approximating those of today. Furthermore, in many places where cuts expose the internal structure of the deposit, the modern surface is seen to truncate the rude bedding of the congelifracts.

RELATION OF SLOPES TO THE TERRACES OF THE AIN

The Ain River breaks out of the Jura Mountains into an ever broadening, flat-floored valley at Neuville about 1 km. west of the site of La Colombière. From this point a series of four terraces has been traced in detail upstream along the Ain and its tributary, the Oignin, as far as the glacial moraine just west of Lake Nantua.[3] Rock-fragment slopes are graded to all but the lowest, or 3–6-meter, terrace. The third highest, 23–28 meters above present stream grade, is known to be related to the outwash plain of the ice which formed the Lake Nantua moraine. At the archeological site a deep excava-

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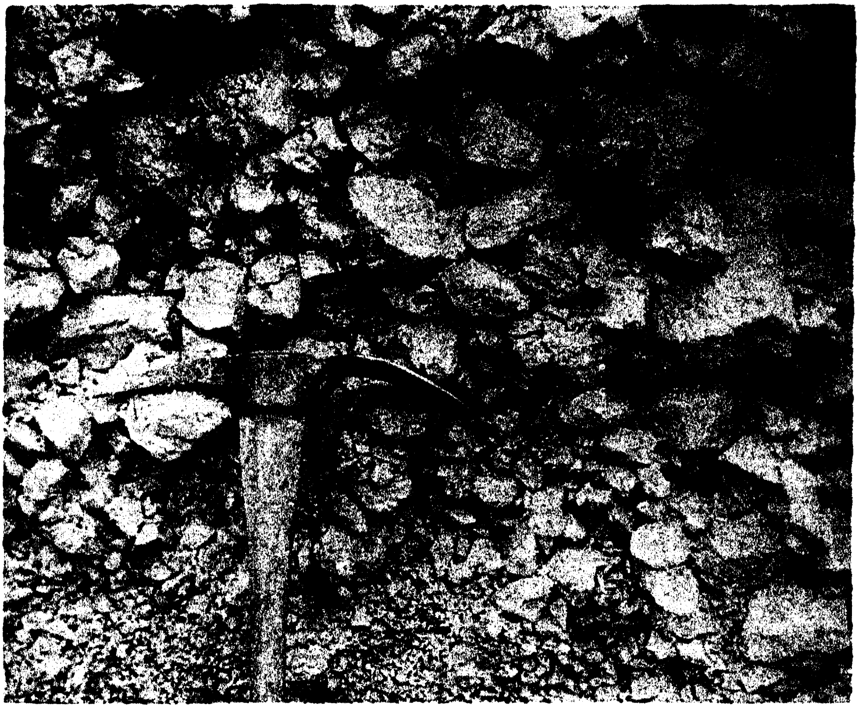
[3] This terrace study and the observations of the relation of the congelifractate deposits to the 23-meter terrace at La Colombière form a part of the geologic study of the archeological site by Bryan and Judson.



"Open-work" texture of rock-fragment slope
at southwest end of Lake Nantua.

tion into this glacial terrace demonstrated an intimate relation between the water-laid terrace deposits and the congelifractate deposits. Angular limestone fragments derived directly from the back wall and roof of the shelter interfinger with the sands and gravel beds within the terraces. Furthermore, beds within the congelifractate slopes on either side of the shelter show continuity with rock-fragment zones contemporaneous with the final stages of terrace deposition. These relationships bear closely on the conditions prevailing during the development of congelifractate slopes.

PROCESS OF SLOPE FORMATION

Rock-fragment slopes are not forming at present in the area studied and are undergoing erosion by processes differing from those which originally formed them. What, then, were the conditions and environment which formed the slopes?

The most convincing evidence that the slopes record a periglacial climate of intense frost action lies in the relation between these slopes and the river-laid deposits of a glacial terrace. Thus, at the site of La Colombière, congelifractate deposits interfinger with glaciofluvial sand and gravel. Here, also, beds within the debris cones are continuous with congelifractate zones involved in the final phases of glaciofluvial deposition. Furthermore, congelifractate slopes along the Ain are graded to two glacial terraces and one periglacial terrace.

The nature of the material comprising the slopes also points to a periglacial origin for the slopes. The angular rock fragments are no longer forming today. They have obviously been derived by rapid and violent mechanical breakdown of local limestone outcrops. Intense frost action is the only effective mechanism which can be invoked. Alternate freezing and thawing during a periglacial or glacial climate would easily account for the production of coarse rock fragments from a dense, somewhat fractured limestone. The downslope movement of these fragments would be accomplished largely by gravity. But the grade of the slopes, even when the effects of modern erosion are considered, is lower than that of slopes produced under simple gravity movement. Rain wash must also have played an important role. Under the periglacial climate which obtained during slope formation, there is no reason to assume a rainfall less than that of today, which is about 40 inches a year. It may very possibly have been somewhat greater. The effectiveness of slope wash is difficult to evaluate. Nevertheless, it is effective today, and there is no reason to suppose that it was not so in the past. Furthermore, if the interior of the slopes were frozen and interstitial voids were ice-filled, remaining so during surface thaws, the slope wash could not escape through the normally porous masses of congelifracts but would be confined to the upper few inches or feet of thawed material. This concentration of water would be most effective in modifying a gravity slope. In addition, it would carry away all but the coarsest material produced during congelifraction. The continuation of this process during the construction of the debris cones would explain not only a grade lower than that of a gravity slope but also the absence of fine material and the consequent presence of openwork texture within the congelifractate slopes.

AGE OF SLOPES

The major climatic prerequisite for the development of the rock-fragment slope is alternate freezing and thawing in a glacial or periglacial climate, as suggested by Cailleux (1948) for his *éboulis* *ordonnés*. As such, it is probable that rock-fragment slopes characterized each advance of the ice during the Pleistocene. The slopes observed, however, are products of the late Pleistocene. Presumably, earlier slopes have long since been destroyed or are present only as fragments buried beneath later accumulations of congelifracts. The debris cones described above are no older than the glacial and periglacial terraces of the Ain, which are considered to date entirely from the last or Würm glaciation. Field work during the summer of 1948 has led Bryan and Judson to the conclusion that the Wurm is multiple in this area. The slopes of the Ain drainage probably owe most of their present expression to the ice advance of Würm III; but many slopes were undoubtedly initiated during the preceding ice advance, or Würm II, the final phases of which are believed to be represented by the moraine at Lake Nantua.

ACKNOWLEDGMENTS.—The writer is indebted to Dr. Hallam L. Movius, Jr., Harvard University, in charge of the archeological excavations near Poncin, Ain, France; to the Viking Fund, New York City, which supported the general investigation of which this note is but a byproduct; and to Professor Kirk Bryan, in charge of the geologic study.

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STONE NETS, STONE STRIPES, AND SOIL STRIPES IN THE WIND RIVER MOUNTAINS, WYOMING[1]

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ABSTRACT

Stone nets, stone stripes, and soil stripes have formed on high, flat erosion surfaces at altitudes ranging from 11,200 to 13,000 feet in the Wind River Mountains, Wyoming. Although the frost-heave theory of Högbom is accepted as accounting for the segregation of coarse material at the surface of ground which has undergone alternate freezing and thawing, certain weaknesses in this theory and in other theories for the origin of stone nets cause them to be unsatisfactory as explanations of those structures in the Wind River Mountains. A theory is developed in which alternate contraction and expansion of frozen ground, under a range of subfreezing temperatures, may produce a net of fractures bounding polygonal areas of frozen ground. Ice wedges form in these fractures and enlarge with successive contractions of the frozen ground. On eventually thawing from the top down, the wedges are gradually replaced by coarse fragments from the upper part of the central polygonal area.

Most of the stone stripes are considered to be the result of solifluction acting on stone nets.

Little definite evidence concerning the age of the soil structures in the Wind River Mountains is available. The stone nets appear to be inactive at present. Solifluction movement is seasonally active in the soil stripes and possibly in some of the stone stripes.

INTRODUCTION

Stone nets, stone stripes, and soil stripes have developed in the mantle rock on extensive high-level erosion surfaces which occur at the northwestern end of the Wind River Mountains of west-central Wyoming. These features are soil structures which occur in regions known to be, or to have been, subject to cold climates. Stone nets are three-dimensional polygonal structures, each composed of a center of fine material bounded by a border of coarse rock debris. Stone stripes are bands of fine rock debris which alternate with channels filled with coarse rock fragments and are oriented parallel to the direction of steepest slope. Soil stripes are similar to stone stripes except that the texture of both coarse and fine material is considerably finer than in stone stripes.

TOPOGRAPHIC SETTING

The Wind River Mountains are characterized by a central divide of high peaks, culminating in Gannett Peak, which has an altitude of 13,787 feet. Flanking the divide at the north end of the range are extensive remnants of a summit erosion surface, which rises toward the divide and is at altitudes of from 11,200 to 13,000 feet. Elsewhere throughout the range, a subsummit erosion surface is well preserved at altitudes ranging from 9,000 to 10,500 feet. Both surfaces are pediment-like in character and, for the most part, are cut on pre-Cambrian crystalline rocks, though tilted Paleozoic strata are beveled by both surfaces at the northwest end of the range. Both surfaces are of Tertiary age.

During Pleistocene time the Wind River Mountains were extensively glaciated, and over forty active glaciers still exist in the shelter of the high peaks along the Continental Divide. The largest of these is 3½ miles long. Most of the remnants of both summit and subsummit erosion surfaces were modified by glacial action. However, at the northwestern end of the range a few fairly extensive remnants of the summit erosion surface project above the upper limits of glaciation, and it is on these remnants that the soil structures (nets and stripes) have formed.

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[1] Published by permission of the Director, United States Geological Survey. Manuscript received October 4, 1948.

RELATION TO GLACIATED AREAS

The soil structures have formed in mantle rock at altitudes of from 11,200 to 13,000 feet. They occur above present permanent snowline, which is at about 10,800 feet, and for the most part above the upper limits of Pleistocene glaciation. Areas in which soil structures occur are indicated in figure 1. These areas are

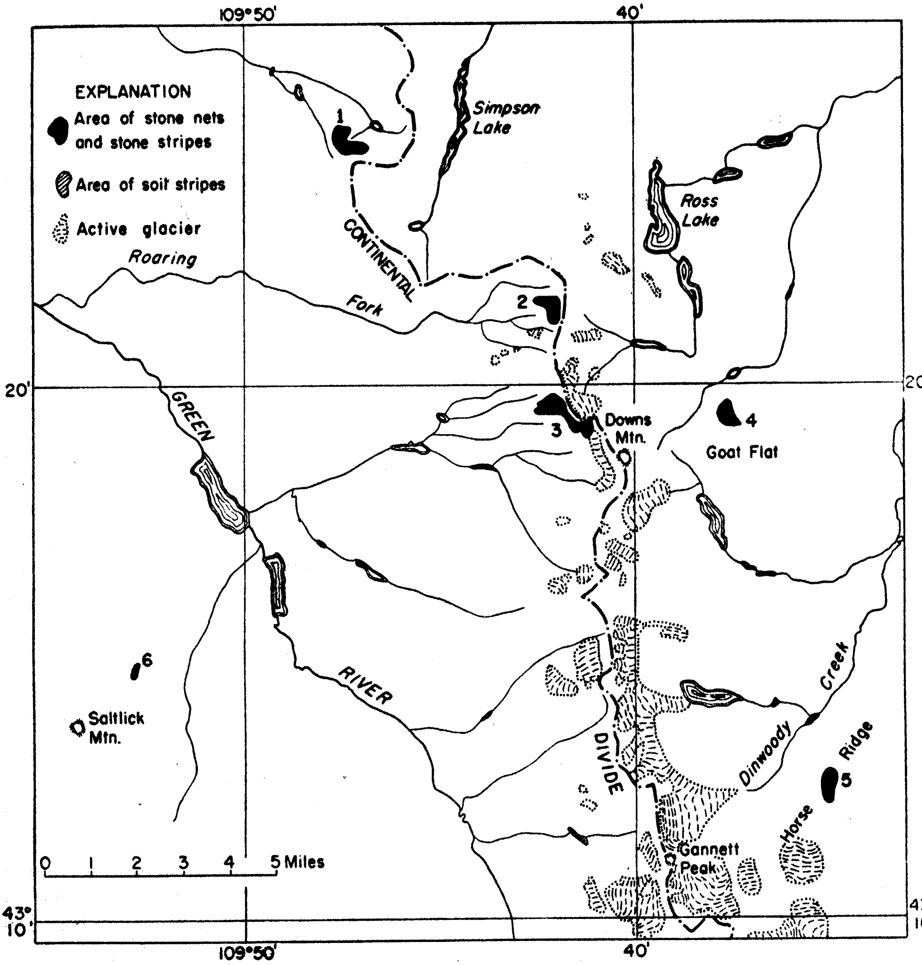


FIG. 1.—Index map of areas of stone nets, stone stripes, and soil stripes at north end of Wind River Mountains, Wyoming.

parts of flat to gently sloping remnants of the summit erosion surface into which cirques have cut headward. They are unglaciated and lack the depressions and irregularities, such as tarns or roches moutonnées, that are common to those parts of the summit surface that have been covered by ice. Instead, they are covered by an extensive mantle rock or felsenmeer composed of sharply angular fragments which exhibit neither the markings nor the shape of typical "glaciated" stones and are derived entirely from the underlying bedrock. Felsenmeer is almost entirely lacking on those parts of the surface that were covered by Pinedale ice and is very much thinner on those parts covered by Bull Lake or Buffalo ice. The terms Buffalo, Bull Lake, and Pinedale refer to the three stages of Pleistocene glaciation in the Wind River Mountains as defined by Blackwelder (1915) and revised as to details by Richmond (1948). Buffalo is the oldest of these stages and Pinedale the youngest.

The area northwest of Downs Mountain (area 3, fig. 1) is an exception to this general picture, in that it is near the head of a shallow basin that is considered to have been a nivation area above the large complex cirque that lies to the west.

RELATION TO BEDROCK

Stone nets and stone stripes are found in areas 1–5 (fig. 1). In area 1, southwest of Simpson Lake, they are developed on unglaciated slopes in felsenmeer derived from the Cambrian Flathead quartzite, a hard, slabby, arkosic quartzite which breaks up into flat, angular fragments and which, on weathering, decomposes into arkosic sand with small amounts of silt and clay.

In areas 3, 4, and 5, stone nets and stone stripes are found on granite gneiss, which is not strongly foliated and which breaks up into chunky angular blocks. The granite gneiss underlying area 2 is strongly banded and consequently breaks up into slabby and elongate fragments. The fine concentrates of the structures in areas 2, 3, 4, and 5 are composed of coarse arkosic sand; silt and clay form a relatively small fraction of the total fines, but a greater fraction than in those structures produced on the Flathead quartzite in area 1.

STONE NETS

The stone nets in the Wind River Mountains are polygonal structures which form a regular netlike arrangement over areas as much as a mile in extent (pl. 1, *A*). They occur on slopes of 4° or less; and, as the slope steepens, they gradually lose their polygonal shape and become elongated until they merge into stone stripes.

The stone nets have from four to six sides and are 5–7 feet in diameter. Each structure consists of a central mass, herein referred to as the "center," which is 2–4 feet across and which contains fine angular gravel, sand, and silt, with a little clay and a scattering of larger rock fragments. Around the center is a channel-like border, 18 inches to 2 feet wide, which is filled with angular rock fragments, 4 inches to 2 feet in diameter. The fragments are loosely packed in an unstable position and in many cases are oriented with their longer axes vertical.

In transverse profile the centers are flat to slightly convex upward. The borders are at essentially the same level as the outer edge of the centers, though many of the rock fragments in the borders project well above the highest part of the centers. In vertical cross section the fine material of the centers becomes coarser downward and, in the three places examined, grades into an unsorted mixture of coarse and fine mantle rock, 1 or 2 feet above bedrock at a depth of approximately 3 feet. The coarse material in the borders in the three places examined extends to a depth of about 2 feet, where it rests with sharp contact on unsorted mantle rock only a few feet above bedrock.

The fine material in the centers of the stone nets is covered by a vegetation mat of grasses and plants common to the alpine zone. The rock fragments in the borders are lichen-covered on their exposed surfaces, but there is no interstitial vegetation.

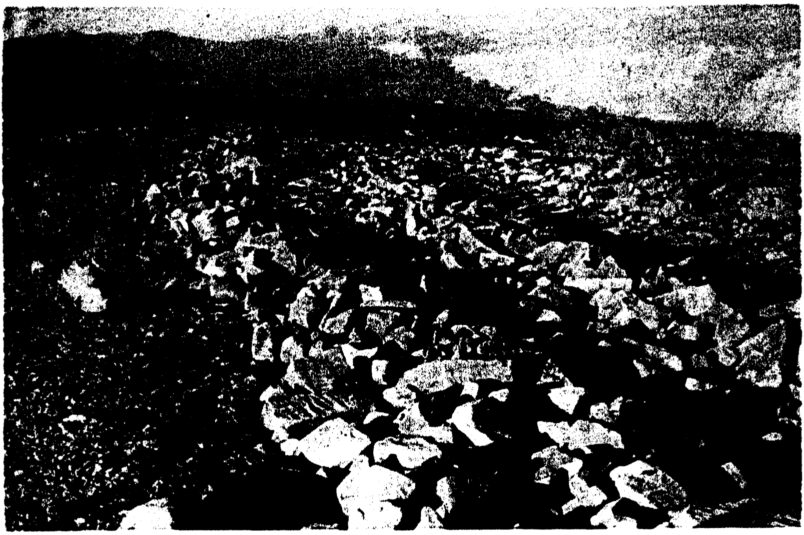
STONE STRIPES

The stone stripes in the Wind River Mountains consist of bands of fine material which alternate with channels filled with coarse rock fragments and are oriented parallel to the direction of steepest slope (pl. 1, *B*). They occur on slopes of 4°–15°. The bands of fine material are from 2 to 4 feet wide, and the material in them is similar in size and shape to that in the centers of the stone nets. The channels of coarse rock fragments are slightly narrower than the adjacent bands of fine material, and the accumulation of fragments is similar in every respect to that in the borders of the stone nets.

In transverse profile the bands of fine material are relatively flat, and the coarse material in the adjacent channels rises 6–12 inches above them. In vertical cross section the material in the bands of fine segregate becomes coarser downward and merges into unsorted mantle rock composed of a mixture of coarse and fine material at depths of from 2 to 3 feet. The coarse fragments in the channels are loosely packed and in an unstable position. Slabby blocks are commonly oriented with their longer axes vertical. The coarse segregate extends to a depth of about 2 feet and rests with sharp contact on unsorted mantle rock a few feet above bedrock. At the bottom of the channels the rock fragments were, in two cases, embedded in sandy material. Water derived from melting snow was observed to be trickling along the bottom of the channels in early July, an indication that they serve as drainage outlets.

Sparse vegetation was observed on some of the bands of fine material, which is loosely compacted and soggy. Lichens cover the outer surface of the coarse fragments in the channels.

On slopes steeper than 15°, stone stripes merge into broad stone streams, which at the surface are composed mostly of angular rock fragments from 6 inches to 2 feet in diameter but, at a depth of about 1 foot, contain considerable material, which ranges in size from fine gravel to silt. Short stone streams occur northwest of Downs Mountain. Washburn (1947, p. 93) has termed similar features "solifluction streams" and interpreted their movement as due to solifluction and creep. Where the slope flattens below an area of stone stripes, these structures do not revert to stone nets. At area 1, southwest of Simpson Lake, stone stripes derived from elongated stone nets encroach over an essentially flat area at the foot of the slope. In places they have merely advanced a few yards and apparently come to rest. In other places, where the supply of material moved down the slopes was greater, they merge into a shallow sheet of rock debris, which spreads out over the flat area. The material at the surface of this sheet is predominantly coarse rock fragments, but fine material fills the interstices at depths of less than 1 foot.



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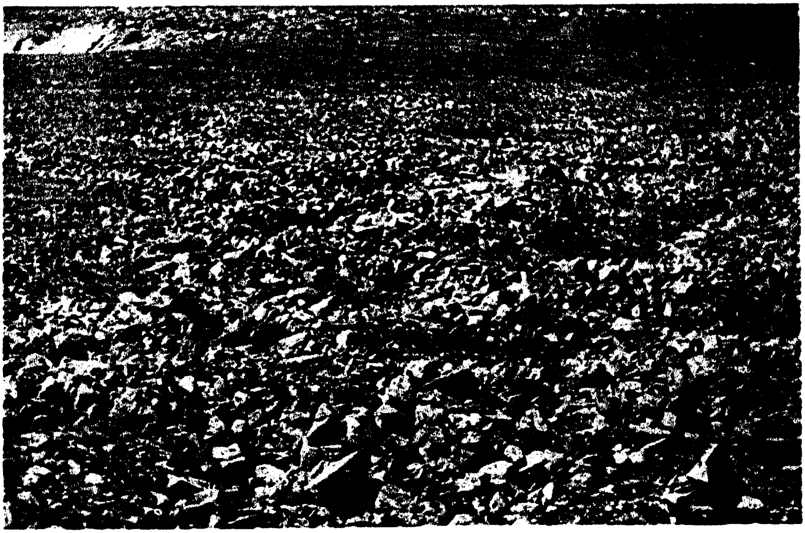

B

A, Stone stripes developed on Flathead sandstone southeast of Simpson Lake (Area 1, fig. 1).

B, Stone nets developed on granite gneiss at head of Roaring Fork (Area 2, fig. 1).

SOIL STRIPES

Soil stripes were found only at area 6, at 11,400 feet on a flat to gently sloping unglaciated divide northeast of Saltlick Mountain. Here soil stripes are formed in arkosic debris derived from a deeply weathered porphyritic granite. They consist of bands of fine sand, silt, and clay which alternate with shallow channels filled with coarse arkosic sand and small rock fragments up to 1 inch in diameter. The bands of fine material are from 2 to 4 inches wide. In vertical cross section they extend to a depth of about 2 inches, where the fine material merges into unsorted soil mantle composed of material ranging in size from clay to gravel up to 1 inch in diameter. The channels filled with coarse material are from 3 to 8 inches wide and extend in vertical cross section to a depth of about 2 inches, where the coarse material rests with sharp contact on unsorted soil mantle. The soil stripes occur on a slope which steepens from 8° to 12°. They are essentially barren of vegetation, though small tufts of grass are scattered over both the bands of coarse material and the bands of fine material.

LOCAL CONDITIONS BEARING ON ORIGIN

Much has been published concerning the origin of polygonal and striped soil structures, and the literature has been ably summarized by Steche (1933) and by Sharp (1942). Local conditions to be accounted for in the Wind River Mountains are the following:

1. The parent material is in most places a felsenmeer composed of angular rock fragments which are, for the most part, the result of splitting and comminution of rock along joints and other fractures.

2. The felsenmeer in which the soil structures have formed is in most places only 3–5 feet thick.

3. The zone occupied by these structures extends nearly to bedrock, and, therefore, at least during the period of maximum activity, there could have been very little unconsolidated frozen ground between the base of the structures and bedrock.

4. The material in the fine concentrates ranges in size from fine gravel to fine sand and silt with little clay.

5. The mineral grains in the fine material occur in about the same proportion and are weathered to about the same degree as those in the underlying bedrock. From this it can be inferred that physical disintegration of the material along joint planes, bedding planes, fractures, intermineral surfaces, and other openings in the rocks has been a more important factor in comminution than has chemical disintegration.

NECESSARY PROPORTION OF FINES

The proportion of fines in the mantle that is necessary for the formation of stone nets or stone stripes is presumably that quantity sufficient to retain enough water to maintain the mass in a saturated condition or at least in a condition that will cause expansion of the mass on freezing. On many of the unglaciated remnants of the summit erosion surface there is an unsegregated felsenmeer composed of angular rock fragments that are mostly greater than 6 inches and up to 4 feet in thickness. The material may be slabby and is invariably loosely packed and in an unstable position. The proportion of fines in this material is very small and is mostly concentrated near the base. It is only where the fine material is abundant enough to fill the interstices between the coarse fragments and to form a layer about 3 feet or more thick that stone nets have formed locally.

EFFECTS OF FROST HEAVE

Where fines have accumulated sufficiently to retain water, frost heave can take place under the proper temperature conditions. Freezing of water in the ground produces layers of clear ice, which, in general, have a cleavage oriented perpendicular to the ground surface. It also may produce ice interstitial to the rock particles. Högbom (1914, pp. 314–315) has shown that, where freezing of a heterogeneous layer of ground occurs, the material, both coarse and fine, is shoved upward more or less uniformly. On thawing, the smaller grains, because of their greater mutual cohesion, are pulled back farther than the coarse fragments, which leaves these fragments in a relatively higher position in the ground.

According to Beskow (1930, p. 635) and others (Hamburg, 1915, pp. 593–594; Hay, 1936, pp. 47–49; Paterson, 1940, p. 120), freezing of water in the ground produces small lenses of ice having a vertical structure under the larger rock fragments and interstitial ice around the fines. As the ground thaws from the top down, the fine grains melt out first because the ice around them is afforded less insulation than that under the coarse fragments. The fines slump in under the coarse fragments, replacing the ice there as it gradually melts. The coarse fragments are therefore prevented from falling back.

Repetition of either process would raise the coarse fragments and eventually cause a concentration of them at the surface, but most authors agree that frost heave and thaw alone can account neither for the borders of stone nets nor for the uniformly spaced three-dimensional polygonal pattern in which they occur.

Högbom (1914, pp. 313–315) originally postulated that frost heave takes place in an outward as well as an upward direction from centers and that the coarse material is thus concentrated between centers to form the borders of stone nets. However, Poser (1933, pp. 105–121) argues that the coarse material may only be moved upward or in a direction perpendicular to the cooling surface.

LOCAL CENTERS OF SILT ACCUMULATION

Many authors have assumed that local centers of silt and/or clay concentration in the mantle retain, through capillary attraction, more water than do adjacent areas of coarser material and therefore become centers of ice heave. This assumption is probably correct, but the origin of a regular pattern of such centers, which is necessary to the production of a regular pattern of stone nets, is not satisfactorily explained. Högbom (1914, pp. 313–315) accounts for stone nets by frost heave taking place from centers, which are presumed to be local concentrations of fines. The cause of these local concentrations and their regularity is not explained. Nansen (1921, pp. 111–112) also believes that local centers of silt and clay concentration are the points of origin of stone nets, and he accounts for them as due to increased comminution caused by a locally increased supply of water accumulating in small surface depressions. The regularity of the pattern is not explained. Elton (1927, pp. 184–185) indicates that the process of comminution may be differential in a vertical plane, where the upper part of the soil is dryer than the lower part, and that, as a result, comminution is locally increased in the lower part. He states: "The polygonal system is thus mapped out underground before it is visible at the surface." He concludes that the fines in the lower part of the soil mantle are heaved up through coarser material in the upper part to produce a hummock of fines surrounded by borders of coarse fragments. The regular arrangement of such silt concentrations is not satisfactorily explained.

These various hypotheses are probably valid, in that where there is a local concentration of silt and clay, a stone ring may form as the result of locally increased frost heave. A group of isolated rings might result from several such centers. The regularity of the pattern of a square mile of adjacent polygonal structures, however, cannot be accounted for by these hypotheses. Why should centers of increased comminution occur in a regular, closely spaced pattern? The concept becomes absurd when the stone nets are small and closely spaced, such as the active miniature nets described by Antevs (1932, pp. 49–50) in the alpine zone of Mount Washington.

Another problem is how frost heave, acting from centers of locally increased comminution, could produce the deep borders of coarse rock fragments observed around the polygonal structures. Elton (1927, p. 187) explains these borders as due to springtime melting of the frozen ground beneath the stones because of the greater insulation of the stones relative to adjacent fine material. This might be a factor in forming shallow gutters at the surface but would hardly be effective in producing border channels 2–3 feet deep.

EFFECT OF SUBFREEZING TEMPERATURE ON FROZEN GROUND

Inactive polygonal soil structures occur only in regions which have been subjected at some time in the past to extremely cold temperatures, and modern active forms are restricted mainly to arctic, subarctic, and alpine zones that have seasonal subfreezing temperatures at present. In temperate zones in many parts of the world, alternate freezing and thawing of the ground take place to varying depths and in some places probably more frequently in the course of a year than in arctic and subarctic regions (Högbom, 1926, p. 254; Elton, 1927, pp. 167–168). Yet polygonal soil structures are not active in temperate zones except under special conditions of drainage and vegetation. It seems plausible, therefore, to relate them to some factor resulting from low temperature.

It is a well-known fact that, although water expands on cooling from 4° to 0° C., ice contracts slightly at temperatures below 0° C. The phenomenon of fractures due to contraction of lake ice may be observed on many frozen ponds on a night when the temperature is below 0° C. Leffingwell (1915, pp. 638–639) and Washburn (1947, pp. 102–103) record that frozen ground cracks with a loud report due to the contraction of the material. These observations were confined to muck and silty or sandy ground. Leffingwell (1915, p. 639) found open cracks as much as 10 mm. wide that crossed frozen ground, and he states that in tundra such cracks commonly divide the muck into polygonal blocks, though locally a crack may cross a flat surface at random.

Washburn (1947, p. 103) states: "With 100 per cent ice a drop in temperature from 32° F. to −40° F. would lead to a linear contraction of one inch in $41\frac{7}{10}$ feet." (This is assuming a linear contraction of 28.3×10^{-6}.) He gives two factors that would cause a fracture once formed to be maintained and to grow in width:

> 1. A crack may become partially filled before expansion [of the adjacent frozen ground] due to rising temperature could close and obliterate the furrow. For instance, wind and gravity may cause infiltration of snow and fines into a newly opened crack. Also water from surface might penetrate a crack extending into frozen ground that was still at a relatively low temperature, and freeze there. 2. A crack once formed would be a line of weakness, as pointed out by Leffingwell [1919, p. 206], and might re-open year after year so that the effect of factor 1 would be additive.

ICE WEDGES ASSOCIATED WITH TUNDRA POLYGONS IN MUCK AND SILT

Washburn (1947, p. 98) found that the borders of mud polygons contained a few thin lenses of clear ice and that stones in these borders were largely sheathed in ice. Leffingwell (1919, pp. 205–212) describes large ice wedges which have a vertical internal structure and which bound polygons in tundra muck in northern Alaska. He observed cracks, up to 5 mm. wide, extending down into the clear ice of the wedge. He shows that frozen ground below the zone of seasonal thaw expands during summer heat and that, in so doing, one of four things may happen: (1) the pressure may melt the ice and close the crack; (2) the material may be sufficiently elastic to absorb the strain, so that no deformation occurs; (3) the formation may deform upward between the bounding ice wedges; or (4) the ice wedges may be deformed.

There is apparently a good chance of the wedge's being preserved until further freezing contracts the ground again and permits additional accretions to it. Leffingwell (1915, p. 642) describes the wedges as thickening at the top with growth and estimates that it would take about three hundred years to develop a wedge 3 feet in diameter. He states that ground-ice wedges were observed only in tundra muck, but he was "unable to say whether ice wedges develop in such deposits (river silts, elevated sand and gravel deposits and soft shales) because only the upper two to three feet of these deposits were exposed which is less than the depth reached by summer thawing."

APPLICATION OF ICE-WEDGE THEORY TO STONE NETS

Leffingwell (1915, p. 639) has observed contraction cracks in frozen muck and silty soil, and Washburn (1947, p. 102) and Paterson (1940, pp. 99–107) have observed them in sand and gravel. These authors agree that thin wedges of ice form in these fractures and may, under the proper conditions, enlarge. Whether a polygonal pattern of wedges will develop would depend on the homogeneity of the material and the flatness of the slope. Ice wedges are known to bound polygons in tundra and in muck or mud. That a polygonal pattern is the normal contraction pattern in homogeneous materials is well illustrated by mud cracks or by columnar jointing in basalt. It should be the normal result of contraction of a relatively flat layer of homogeneous frozen ground in which cooling took place perpendicular to the ground surface. The width of the fractures will depend on the amount of change in temperature and the coefficient of expansion of the frozen material. The amount of expansion of ground on freezing approximates a direct ratio to the volume of ice formed in it. Consequently, the coefficient of expansion of frozen ground may be considered to approximate that of clear ice. Assuming the coefficient of linear expansion of clear ice as between 52.7×10^{-6} at 0° C. and 50.5×10^{-6} at −30° C., which figures are quoted by Dorsey (1940, p. 473) from Jacob and Erk (1928) as being the most satisfactory, it should be possible with a drop in ground temperature of 30° C. to obtain a series of fractures about 1.5 mm. wide around a prospective stone net, 1 meter in diameter. The width of the fractures, however small, is large enough for the penetration of moisture and the formation of an ice wedge, which would tend to protect the fracture from being destroyed on expansion due to rising temperature and would allow it to be enlarged by about 0.05 mm. with every degree centigrade fall in temperature. Alternate rising and falling of temperature, with a total range entirely below 0° C., might occur many times in the course of an arctic or Pleistocene winter with no more than surficial melting of the ground or snow before the summer thaw set in. If the basis for this hypothesis is sound, wedges of clear ice would eventually form in a regular polygonal pattern, which can be accounted for as a normal fracture pattern brought about by contraction of a layer of homogeneous material perpendicular to the cooling surface. During rises in temperature that did not thaw the ground to any significant depth, the material in the centers would expand upward, and coarse fragments on the surface would become loosened and tend to slide into furrows overlying the wedges. During complete or seasonally deep thaws the ice wedges would gradually melt, at least to the depth of effective thaw, which for the areas of stone nets in the Wind River Mountains was probably the full thickness of the mantle, or a depth of about 3 feet. The coarse material on the surface would settle more deeply into the fractures as the clear ice melted, and both coarse and fine material would slump in from the walls. Filling of the wedges in this manner would account for the sharp contact between the coarse material in the channel and the unsorted mantle rock beneath. There would be a tendency for the fines to be washed down and, slope permitting, out. Because of their greater mutual cohesion, the fines would also tend to be reincorporated in the central mass more easily than would the coarse fragments. Any coarse material in the centers would eventually be brought to the surface by alternate frost heave and thaw, even though thawing may have been infrequent in the course of a year. By this process a repetition of alternate changes in temperature, with most of the changes confined to fluctuations below 0° C., could have caused the development of the mature stone nets observed in the Wind River Mountains.

ORIGIN OF STONE STRIPES

The formation of stone stripes on slopes of from 4° to 15° by frost action is difficult to understand. Alternate frost heave and thaw would bring the coarse fragments to the surface, provided that movement of material down the slope during thaws did not obliterate this process. Once stripes have formed, their propagation down the slope by solifluction is readily agreed upon; but what initiates segregation of the material into coarse and fine bands has not been satisfactorily explained. The theory of local silt concentrations is no more plausible for this situation than it is as a factor in the initiation of stone nets. It does not seem any more plausible that contraction of the ground when frozen would produce a polygonal fracture pattern or that this would be preserved, even if it contained wedge ice, when thawing and downslope motion of the material set in. Paterson (1940, p. 100) indicates that frost cracks form on gentle slopes but that they are oriented at random. Should there be a tendency for them to be oriented down the slope, it would be easy to see how they would be favored for development into stripes by mass-wasting through either solifluction or gravity.

A more probable explanation is that segregation of the material takes place as stone nets on a relatively flat surface and that these structures migrate through solifluction from their point of origin down the slope to their present position, first, as elongate stone nets or stone garlands and, finally, as stone stripes. Such a progression of structures occurs at area 1, southwest of Simpson Lake. The extent of their migration can be considered an indication of the extent and volume of material involved in mass-wasting on the slope.

ACTIVITY, PRESENT AND PAST

Probably only the soil stripes are active at the present time, though some solifluction movement of the stone stripes may be going on. Formation of stone nets has apparently ceased. Some frost heave may still occur in the centers of the polygonal structures, though the vegetation mat is little disturbed. No movement of the coarse material in the borders is indicated by the fact that the outer surfaces of the rock fragments are covered by lichen and that none of these fragments has been disturbed sufficiently to expose the clean undersides.

Because of the location of most of the areas of stone nets and stone stripes above the limits of glaciation, it is difficult to determine the time at which they began to form. Though they are inactive at present and no ice or permanently frozen ground is associated with them, they may be modern in the sense that they formed after the Pinedale stage of glaciation during and since the recent readvance of the ice when conditions necessary to their formation probably prevailed. The structures at area 3 have formed since the Bull Lake glacial stage, inasmuch as this area was a Bull Lake nivation area. The structures in the other areas could have formed as early as the Bull Lake or even the Buffalo stage. However, regardless of when they began to form, it is likely that they lay dormant during interstadial periods and were rejuvenated and further developed during each successive cooling of the climate and re-advance of the ice to the present.

SUMMARY

Stone nets, stone stripes, and soil stripes have formed in the Wind River Mountains on high, flat erosion surfaces above the upper limits of glaciation. The stone nets and stone stripes formed in a relatively thin felsenmeer consisting of coarse and fine rock fragments, sand, and silt, with relatively little clay. In general, the finer material settled into the lower part of the felsenmeer.

Frost heave and thaw, though effective in segregating coarse material from fine by raising the coarse material to the surface, could not alone account for the formation and regular pattern of polygonal and striped structures over areas as much as a mile in extent. The theory of local centers of accumulation of silt and clay in the ground is discarded as a possible initiating factor for the origin of the pattern of stone nets in this area. Because the distribution of these structures is confined to areas which have been subject to subfreezing temperatures, it is logical to relate the origin of the structures to low temperatures. Ice is known to contract at temperatures below 0° C. Observations by others indicate that frozen ground cracks at low temperatures and that ice forms in such contraction cracks and may develop into ice wedges, which—at least in the case of tundra muck—have been observed to bound polygonal structures. Polygonal structures are the common result of contraction of a layer of homogeneous material perpendicular to the cooling surface, as is well illustrated by columnar jointing of basalt or in the formation of mud cracks. A regular pattern of polygonal fractures could form in a relatively flat layer of frozen ground as a result of contraction of the ground during periods of subfreezing temperature. Ice wedges could form through a repetition of alternate changes in temperature, with most of the changes confined to fluctuation below 0° C. Filling of these wedges with coarse material from the surface during seasonal complete thaws accounts for the formation of border channels of coarse material around the centers of the polygons and the sharp contact between this coarse material and the unsorted mantle at the bottom of the channels. The concentrate of fine material in the center of the polygonal structure is the result of the raising of the coarse fragments to the surface by frost heave and thaw. The stone stripes and soil stripes are thought to be the result of elongation and separation of stone nets into stripes through solifluction rather than to have been formed in place on the slope. Solifluction movement of the soil stripes and some of the stone stripes is the only activity of these structures going on at present.

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SOME PERIGLACIAL FEATURES IN CENTRAL MONTANA[1]

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ABSTRACT

Two types of structures represent the results of former intensive frost action in central Montana. Involutions, which consist of local deformation and interpenetration in stratified materials, are common throughout the area and occur in a variety of forms. They are believed to have developed by differential freezing and growth of ground-ice, in a seasonally thawed zone overlying perennially frozen ground. Wedge structures, found at a few localities, consist of wedge-shaped vertical fissures in weathered bedrock. The fissures form polygonal networks and have been filled by slumping of surface materials. They are believed to have been opened by ice veins developed when the ground became perennially frozen. A mean annual temperature at least 8° C. below that of the present would be required for the development of perennially frozen ground. This period of colder climate probably occurred at the time of the last glaciation of adjoining areas, which may belong to the Middle Wisconsin substage.

INTRODUCTION

A considerable literature exists dealing with surficial structures produced by frost action.[2] Some of these structures are being formed in high latitudes and at high altitudes, where the modern climate promotes intensive frost action. Others have been found in regions where the climate does not permit their formation at the present time. The features here described belong to this latter group of "fossil" forms, produced under a previously existing climate more rigorous than that of the present. That former climate is commonly referred to as "periglacial," because of its association with near-by glaciation. The periglacial environment, according to Sharp (1942*a*, p. 115), is characterized by "low temperatures, strong winds, and many fluctuations across the freezing point at certain seasons."

GEOLOGICAL SETTING

The structures to be described occur in central Montana, between Lewistown and the Rocky Mountains (fig. 1). This area lies in the western part of the Great Plains, from which, in this vicinity, rise several mountain ranges lying to the east of the main ranges of the Rockies. The bedrock at all the localities described is of Cretaceous age: Two Medicine formation (Montana group) at locality 4; Colorado shale at localities 1, 2, 3, 7, 8, 9, and 10; Kootenai formation at localities 5 and 6. At many of the localities the structures occur in terrace gravels and other sediments. These gravels have been mapped by Alden (1932, pl. 1) as belonging to his No. 2 and No. 3 benches or stream terraces. The No. 2 bench he attributed to the "pre-Iowan interglacial stage (possibly pre-Kansan)," and the No. 3 bench to the "pre-Wisconsin interglacial stage," possibly including some Wisconsin glacial outwash. The gravels are derived principally from the mountains to the south and west. Two-thirds of the pebbles and cobbles consist of several varieties of limestone of Mississippian age, and the other third consists mainly of sandstone, quartzitic sandstone, and shale of Cretaceous age.

At least twice during the Pleistocene epoch a lobe of the Keewatin ice sheet extended southward between the Bear-

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[1] Manuscript received November 27, 1948.

[2] For a brief historical summary see Sharp (1942*a*, p. 114). Other general references are Sharpe (1938, pp. 33–46), Sharp (1942*b*), Troll (1944), and Zeuner (1945, pp. 6–13).

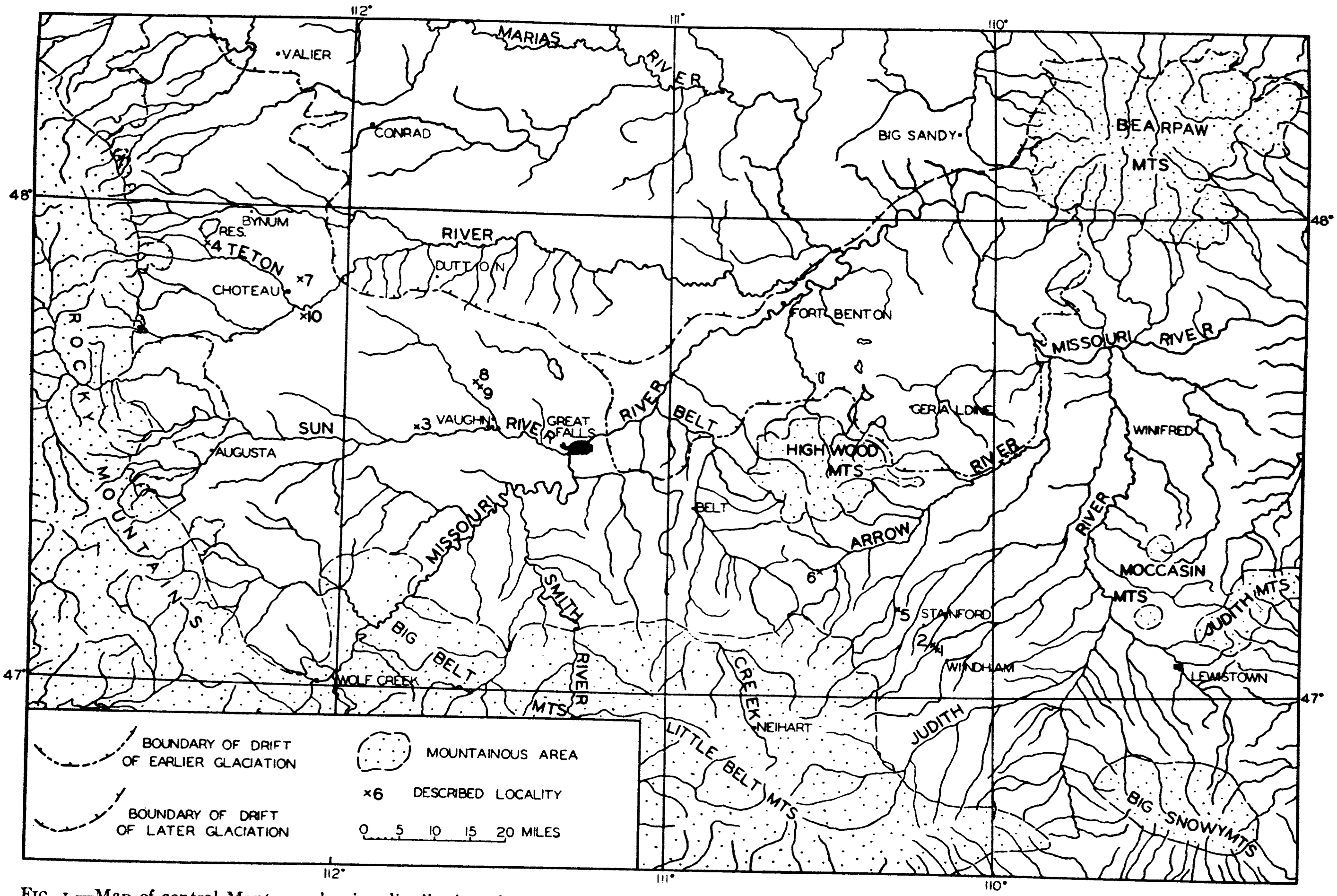

FIG. 1.—Map of central Montana, showing distribution of mountainous areas and drift sheets (after Alden, 1932, pl. 1) and positions of described localities

paw and Rocky Mountains nearly to the site of Great Falls. The age of these advances is as yet uncertain, but the earlier is believed by Alden (1932, pp. 85–88) to be of Iowan or Illinoian age. The later advance he believes to be of Early or, in modern terms, Middle, Wisconsin age or younger than the Iowan substage and older than the Altamont moraine (Mankato substage) (Alden, 1932, pp. 93–96). The ice of the earlier advance crossed the valley of the Missouri River northeast of Great Falls, damming the stream and producing a proglacial lake of considerable extent (Calhoun, 1906, pp. 30–31).

Small ice tongues issued from the major valleys of the Rockies onto the western edge of the Great Plains. These may presumably be referred to the Middle Wisconsin substage because of the freshness of the topography of their moraines. Within the limits of area of the map (fig. 1), so far as is known, none of the small mountain groups east of the main ranges supported mountain glaciers.[3] However, evidence of mountain glaciation has been found in the Castle Mountains, south of the Little Belt Mountains (Weed, 1899, p. 7), and in the southern part of the Big Belt Mountains (Pardee, 1925, pp. 37–38).

DESCRIPTION OF INVOLUTIONS

The term "involution" has been used by Denny (1936) and by Sharp (1942*a*) for local deformation and interpenetration produced in stratified materials by frost action. Many road cuts and some stream banks in the area display involutions. Four exposures at which these features are particularly clearly developed and well exposed have been chosen for detailed description.

LOCALITY 1

Locality 1 is at a road cut on the northeast side of U.S. Highway 87, about 2.2 miles northwest from Windham (highway distance). The materials in this cut (pl. 1, *A*) are terrace deposits on Alden's No. 2 bench and range in grain size from gravel to fine silt or clay. The underlying Colorado shale is not exposed here but may lie at a depth of about 5–20 feet below the base of the cut. At the place where the man's figure may be seen in plate 1, *A*, the slumped material was excavated, uncovering the structures shown in plate 1, *B*, and figure 2. The material exposed in the lower part of the excavation is pebble gravel, somewhat cemented by calcium carbonate. Above this is a thickness of about 2½ feet of stratified sand and silt, white to tan and orange in color. Overlying these beds is an ancient immature soil zone about 3½ feet thick, consisting of a light-colored "B" horizon enriched in calcium carbonate, and a dark-gray or black "A" horizon containing a considerable amount of organic matter.

All these materials are strongly involuted. The deformation of the stratified materials seems to have involved movements that were largely vertical, as indicated by upward intrusion of masses from the lower layers into those above. The materials of the ancient soil show a series of structures resembling drag folds, overturned toward the right side of the photograph. This suggests that slight movement to the right (southeast) occurred in the materials near the surface. This is the direction of surface slope, which amounts to about 3°. The involuted nature of the A horizon of the ancient soil is clearly shown in plate 1, *A*. At the time of development of the involutions, the B soil horizon was brought to the surface at some places (as at the left side of plate 1, *B*). Since that time, a new carbonaceous A horizon, less than 6 inches thick, has developed across all the surface materials, including the much thicker A horizon of the ancient soil. This modern soil does not clearly show a B horizon enriched in calcium carbonate. When the face of the excavation is cut back, it is found that most of the involutions do not persist more than a very short distance in a direction normal to the cut face and therefore consist of plugs or narrow tongues rather than folds of considerable lateral extent. This is particularly true of the structures in the lower part of the section, beneath the deformed soil zone. Most of these are subequidimensional in plan.

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[3] Weed (1899, p. 7) states: "No evidences of glaciers are found in the Little Belt Range." However, the topographic map of the Little Belt Mountains Quadrangle shows on the east side of Baldy Mountain (9,000 feet in altitude) in the northern part of the range two valleys of cirque-like form, one of which contains a small lake.

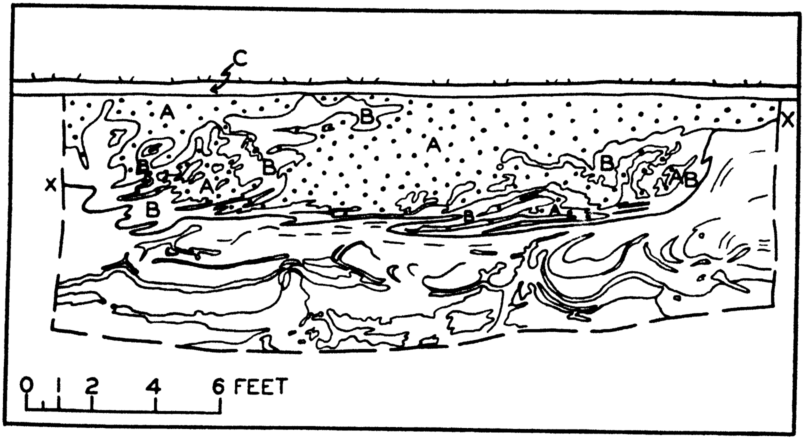


Fig. 2.—Diagram of the involutions at locality 1 shown in pl. 1, *B*. Symbols: *A*, the A horizon of the ancient soil (*dotted pattern*); *B*, the B horizon of the ancient soil; *C*, the A horizon of the modern soil. The heavy line *X–X* represents the upper contact of the underlying terrace deposits, consisting of silt, sand, and gravel. The heavy broken line represents the limits of excavation.

LOCALITY 2

Locality 2 is at a road cut on the northeast side of U.S. Highway 87, about 3.0 miles southeast from Stanford and 1 mile northwest from locality 1 (highway distances). The materials in this cut are terrace deposits and soil zones, whose relations are similar to those at locality 1 and which also occur on a segment of the No. 2 bench. The ancient soil zone is here between 2 and 3 feet thick, and the modern soil is less than 6 inches thick. The calcareous terrace sands and silts are white to orange and reddish-brown in color. The cut displays involutions like those at locality 1, and, in addition, there are several vertical pipelike masses of terrace sand and silt extending upward nearly or quite to the base of the modern soil zone. The ground surface is essentially flat, and there is no evidence of significant lateral movement of the surficial materials.

LOCALITY 3

Locality 3 is at a road cut on U.S. Highway 89, about 24 miles west from Great Falls and 3.6 miles west from the intersection with Montana Highway 20 (highway distances). Good exposures are provided on both sides of the highway and in a gravel pit immediately south of the road cut. This locality is at the eastern edge of a segment of the No. 2 bench. The terrace deposits exposed consist of pebble and cobble gravel, with scattered boulders and sand lenses. As at the other localities, there is a modern carbonaceous soil horizon less than 6 inches thick. The ancient deformed soil, which is so conspicuous at localities 1 and 2, is not present, unless some of the uppermost highly calcareous material represents part of the B horizon of that soil.

A conspicuous feature of the cut on the north side of the highway is the involution shown in plate 1, *C*, consisting of a tongue of sandy material extending 5 feet downward into gravel. The form of this involution, as of some of those at locality 1, suggests overturning toward the right. This is the downslope direction, but the angle of surface slope is very slight: 1°–2° or less.

Plate 2, *A*, shows an exposure in the pit south of the road cut. Pointed tongues of gravel extend into the overlying sand and sandy silt. In undisturbed gravel most of the stones tend to lie with their long axes nearly horizontal or somewhat imbricated; but in distorted gravel, such as that shown in the figure, the long axes of the stones tend to lie parallel to the contact between the beds. At the tips of the gravel tongues the long axes of the stones are approximately vertical. These involutions may be called "festoons," from their resemblance to loops like those formed by a limp cord suspended from a series of points along a taut wire. Festoons are also found in relatively homogeneous gravels which lack a gravel-sand contact to show the deformed character of the materials. In such cases the form of the festoons is defined by the orientation of the long axes of the stones. An example of this situation is found in a small gravel pit at locality 7, about 2.2 miles north-northeast of Choteau, near the southeast corner of Section 7, T. 24 N., R. 4 W. Involutions quite similar in form to these festoons are found in weathered, thin-bedded sandstone of the Colorado formation at the north end of the road cut at locality 9, on U.S. Highway 91, about 5.6 miles north of its junction with U.S. Highway 89 just east of Vaughn (highway distance).

PLATE 1

A, View to northeast at locality 1, showing prominent involutions in ancient soil zone. Segments of No. 2 bench in background, and Judith Mountains in distance at right.

B, Involutions in ancient soil zone and terrace deposits at locality 1 (excavation where man's figure may be seen in pl. 1, *A*). For diagram see fig. 2.

C, Involution in sand and gravel of No. 2 bench at locality 3.

PLATE 2

A, Festoons in gravel of No. 2 bench at locality 3. Note tendency toward vertical orientation of pebbles in tongues of gravel.

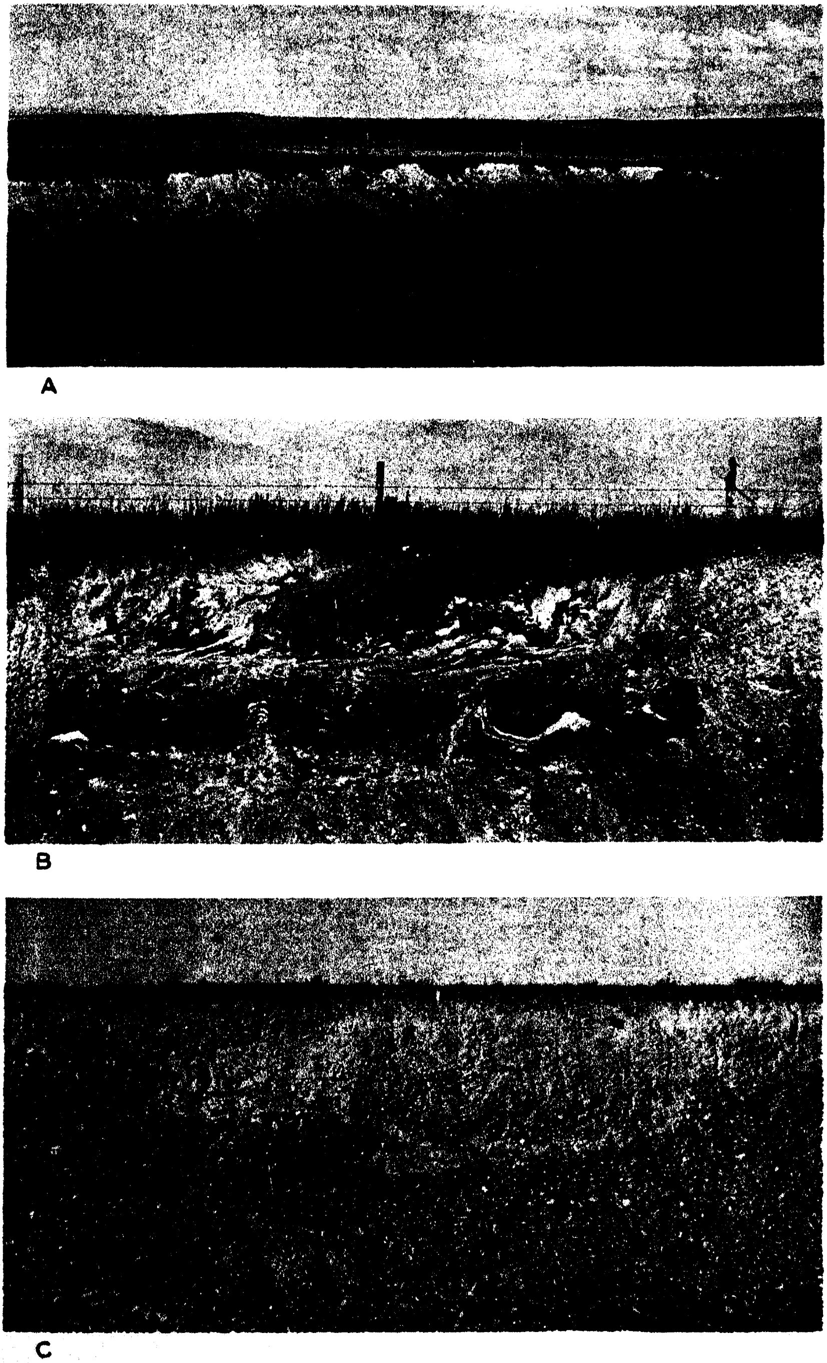
B, Involutions in terrace deposits and shale at locality 4. Bank is about 7 feet high. For diagram see fig. 3.

PLATE 3

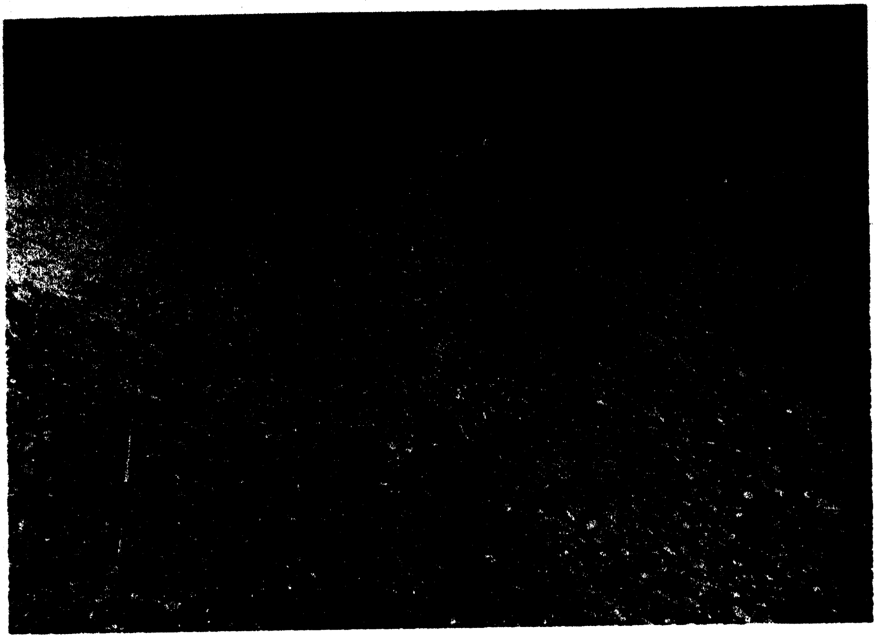
A, View to northeast at locality 9, showing wedge structures (*dark bands*) arranged in a roughly polygonal network. Large wedge at right extends to depth of 13 feet. Broad flat upland in background.

B, Wedge structures in shale at locality 8. Bank is about 7 feet high. For diagram see fig. 7.

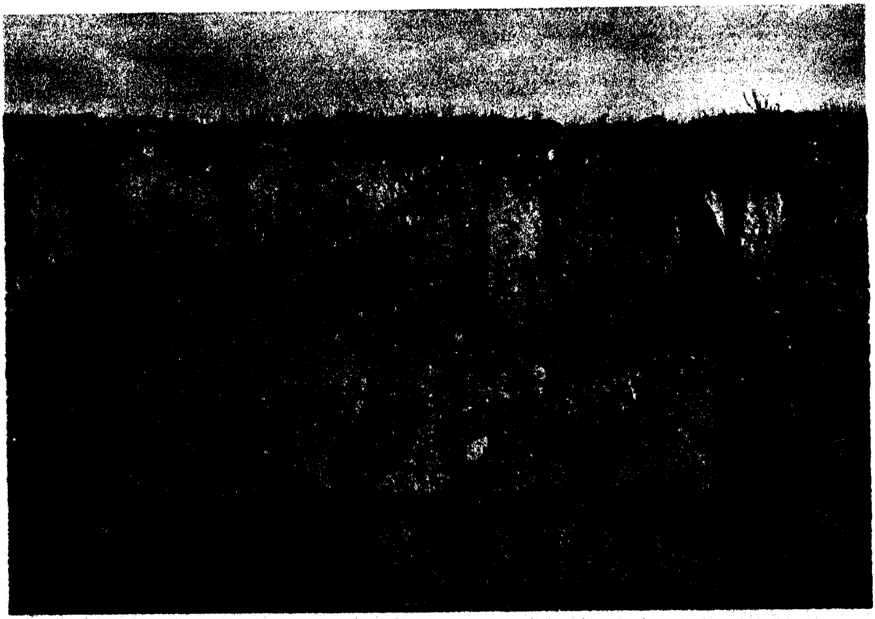
C, Wedge structures in shale at locality 8. For diagram see fig. 8.



Involutions

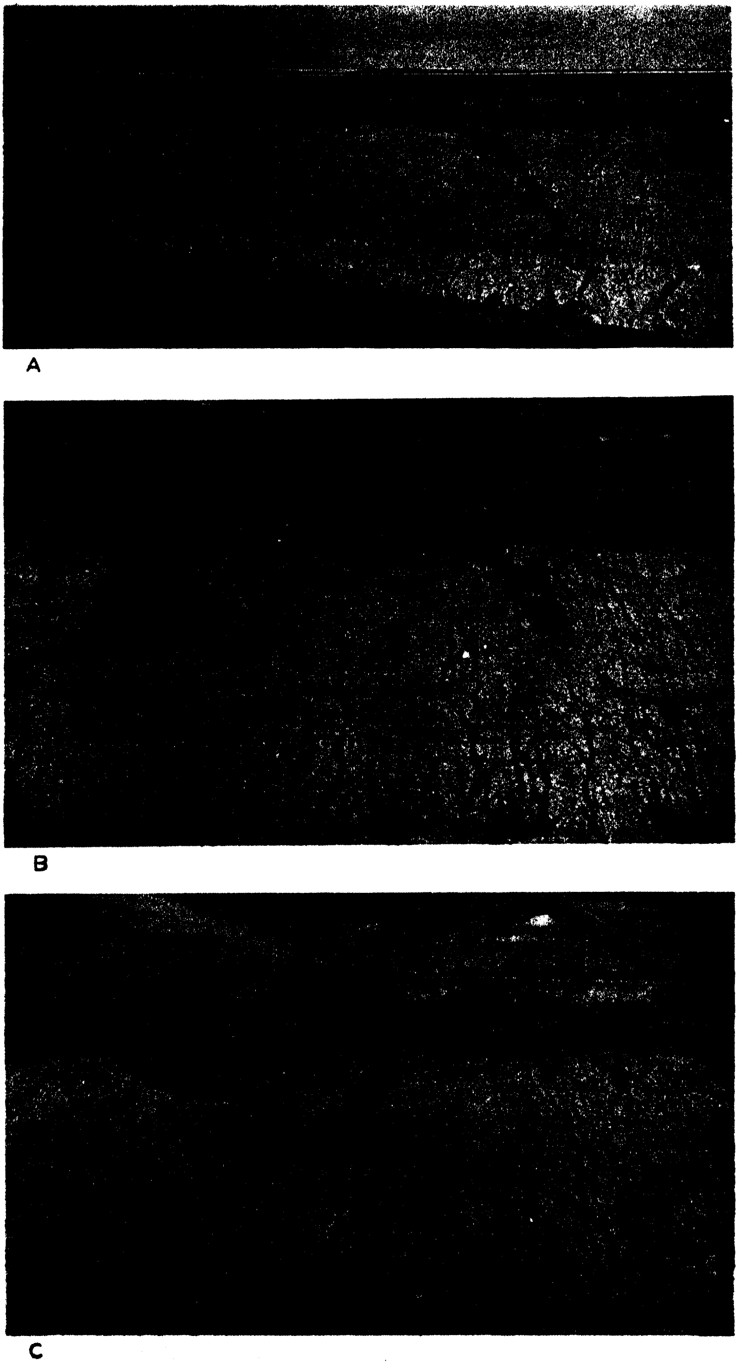


A



B

Festoons and involutions



Wedge structures

LOCALITY 4

Locality 4 is about 13.5 miles northwest of Choteau and 1 mile south of Bynum Reservoir, in the northeast quarter of Section 24, T. 25 N., R. 7 W. The exposure is in the southeast bank of a gully cut by water diverted from the Teton River to Bynum Reservoir. The locality lies at the northern edge of a gravel-floored abandoned valley of the Teton River; and for a short distance along the gully the contact of the feather edge of the gravel with bedrock is shown. This gravel area is not shown by Alden on his geologic map (1932, pl. 1), but it may be traced eastward to its confluence with the broad area of gravel between Bynum and Choteau, which he attributed to his No. 3 or pre-Wisconsin bench level (1932, p. 60). However, it may also be traced westward, where its relations to the Teton River and to the moraines of the Teton River glacier indicate that this gravel was probably deposited during a halt in the early part of the recession of that glacier.

The bedrock in the exposure is green shale of the Two Medicine formation, the top several feet of which are soft and crumbly because of weathering. The terrace deposits above the shale in plate 2, *B*, and figure 3 consist of 2–3 feet of interbedded sand and gravel, overlain by 3–4 feet of sandy silt.

This entire sequence of materials is involuted, as is shown in plate 2, *B*, and figure 3. Figure 4 represents the same bank a few feet to the east, where the gravel is thicker and rests directly on bedrock. Seen in vertical section, some of these structures resemble those produced by channeling or cut-and-fill action. However, they lack the linear continuity in the third dimension which such features possess, and the extension of the tongue of gravel several feet upward into silt in figure 3 cannot be explained on this basis. The form of some of these structures superficially resembles that of filled potholes. This mode of origin is obviously inadequate to explain the gravel-sand relationship shown in figure 3. Furthermore, close inspection of the shale in the lower part of the bank shown in figure 4 reveals that deformation concordant with the subspheroidal masses of gravel extends to a depth of several feet in the weathered bedrock. The small gravel masses which, in figure 4, appear to be isolated from the main body of

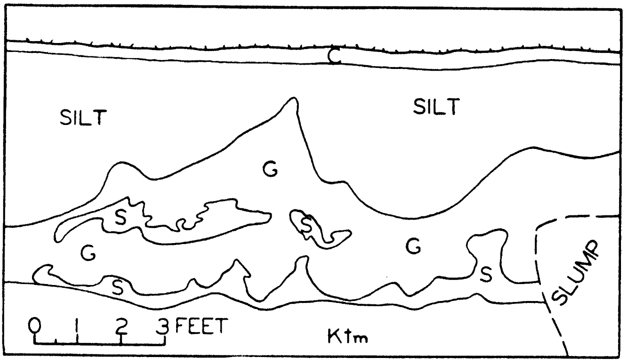


FIG. 3.—Diagram of the involutions at locality 4 shown in pl. 2, *B*. Symbols: *C*, the humus-rich A horizon of the modern soil; *G*, gravel; *S*, sand; *Ktm*, shale of Two Medicine formation.

gravel are perhaps connected with it in a direction at right angles to the plane of the section.

ORIGIN OF INVOLUTIONS

INTRODUCTORY STATEMENT

The involutions are secondary structures, superimposed on the surface materials after their deposition or formation. This is particularly clearly shown by the deformation of horizons of a soil zone or by the interpenetration of materials of different ages, such as soil zones, terrace deposits, and Cretaceous shales. Six processes[4] are potentially capable of producing structures like the Montana involutions. These are: (1) ice shove, (2) volume change in bentonite, (3) differential loading, (4) mass movement, (5) tectonic deformation, and (6) intensive frost action.

ICE SHOVE

Overriding of unconsolidated materials by glacial ice could not have produced the involutions, since all the localities described lie outside the boundaries of glaciation. The grounding of icebergs or floe ice in shallow water must be considered with regard to locality 3, since the area in which it lies was thought by Calhoun (1906, p. 30) to have once been submerged in a proglacial lake. However, such characteristics as the festoons of oriented pebbles and the lack of linear trend would not be expected of structures produced by dragging ice. Another objection may be made on the basis of time relations. The proglacial lake existed only during the earlier—Iowan or Illinoian—stage of glaciation. The modern soil zone, which has been formed since the involutions were formed, is of essentially the same thickness at all localities observed, so that the involutions are probably of the same age. This soil zone is so thin, however, that the involutions, including those at locality 3, are probably no older than the later or Middle Wisconsin glaciation and were formed long after the proglacial lake had ceased to exist.

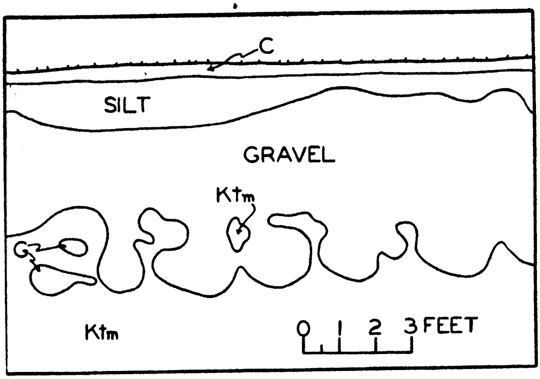


FIG. 4.—Diagram of involutions at locality 4, a few feet east of those shown in pl. 2, *B*, and fig. 3. Symbols as in fig. 3.

VOLUME CHANGE IN BENTONITE

Considerable quantities of bentonite occur in the Cretaceous rocks within the area studied, and bentonitic beds are actually present in some exposures of involuted bedrock. It seems certain, however, that none of the involutions is the result of deformation caused by change in volume of the bentonite. The distribution and forms of involutions bear no systematic relation to bentonitic zones, and involutions are found in some exposures of bedrock containing little, if any, bentonite. A road cut a few hundred feet west of that described as locality 2 displays a thickness of about 15 feet of terrace deposits overlying the Colorado shale. Involutions occur near the surface but die out downward, so that none is shown in the lower part of the terrace deposits or in the 15 feet of shale exposed. Numerous other exposures demonstrate that involutions not only may show no relation to the type of bedrock but also may be confined to materials overlying the bedrock. Anhydrite, like bentonite, undergoes considerable change in volume by taking up water but is not present in this area in significant quantities.

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[4] Several others are listed by Sharp (1942*a*, p. 126), but they are so unlikely as not to require discussion.

DIFFERENTIAL LOADING

Sharp (1942*a*, p. 127) discusses briefly the deformation of relatively fine-grained, water-saturated materials that may occur under a differential load. A contact between sand or gravel and underlying silt or clay is particularly susceptible to such disturbance. However, the involutions were formed under little or no overburden, as a soil zone is involved, and it seems quite unlikely that any load was applied on this soil and subsequently removed. This hypothesis is particularly inapplicable to the coarse sediments of locality 3.

MASS MOVEMENT

It is possible that differential movements within masses of weak materials undergoing landsliding might produce structures similar to some of the involutions. None of those described is believed to have been formed in this manner, and the forms of most of the involutions indicate vertical rather than lateral movement. Involutions were observed at numerous localities, such as locality 4, where the topographic and geologic relations of the involuted materials are highly unfavorable for the occurrence of mass movement.

At locality 5, a road cut more than 20 feet deep on the north side of U.S. Highway 87, about 4.2 miles west of Stanford (highway distance), occurs a series of terrace sands and gravels overlying beds of the Kootenai formation. The terrace deposits dip eastward at an angle which increases downward in the section from 30° to 60°, and the Kootenai strata are nearly vertical. The locality lies on the east side of a creek valley, and the structure described is believed to be due to landsliding, of the type called "slump" by Sharpe (1938, p. 65), on the wall of that valley. Definite landslide topographic forms (called to the writer's attention by Andrew G. Alpha) exist along the valley wall a short distance to the south and to the north. Associated with the major structure are small-scale folding and crumpling of the strata; these minor structures bear little resemblance to involutions. The modern soil zone, formed on this landslide block since the landsliding took place, is of about the same thickness as that formed on involuted materials. It seems likely that the same conditions that promoted the development of involutions may also have promoted the landsliding, which is, like the involutions, a fossil phenomenon at this locality (cf. Smith, 1936).

Deformation of sediments contemporaneously with deposition generally takes place as subaqueous sliding under the influence of gravity. This mode of origin cannot be applied to involutions affecting soil zones or to involutions whose form suggests vertical rather than horizontal movements of materials or to involutions affecting both surficial sediments and the underlying bedrock.

TECTONIC DEFORMATION

Intricate deformation of soil zones and terrace deposits cannot be attributed to tectonic movements; but at several localities involution-like structures seem to be confined to bedrock. At locality 6, a road cut on the south side of U.S. Highway 87, about 2.2 miles west of Geyser (highway distance), is exposed a small overturned fold with an amplitude of about 8 feet, in shale of the Kootenai formation. Several feet of terrace deposits overlying the shale are not affected by the folding, and it is believed that this fold is of tectonic origin. Other exposures were observed at which the available evidence did not show clearly whether the structures were of tectonic origin or were involutions.

INTENSIVE FROST ACTION

General statement.—The undisturbed character of the modern soil zone shows that the involutions are not being formed at present, and no agent adequate to produce these structures is now available. They are readily explicable, however, as the results of frost action under the influence of the colder climate which must have existed in this region when the margin of a Keewatin ice sheet lay but a few miles to the north. The surficial structures resulting from contemporary frost action in arctic regions have been observed and described by numerous authors (see references in n. 2). The formation of these structures takes place principally in a thin, periodically thawed layer overlying perennially frozen ground and is assisted by the segregation of masses of ground-ice. These conditions may have existed in a belt peripheral to the ice sheet of any glacial stage or substage. The forms of the involutions suggest that the materials had a rather high degree of mobility at the time that deformation occurred. The considerable differences among the forms of the involutions at the four localities described may be attributed to differences in the textures and lithologies of the materials, and possibly also to differences in groundwater supply and in exposure to climatic factors. Structures superficially similar to involutions but of entirely different origin also occur within the area, as is noted above in the sections on mass movement and tectonic deformation.

Role of ground-ice.—The knowledge of the mechanics of frost heaving is derived principally from the experiments and observations of Taber (1929, 1930, 1943; see also Muller, 1947). The expansion of freezing interstitial water in porous materials is quantitatively insufficient to produce excessive frost heaving, which is caused by the segregation of horizontal layers and lenses or vertical, wedgelike veins of ground-ice. Within inhomogeneous materials differential freezing may take place, resulting in differential segregation of ice and differential heaving as water is drawn to growing ice masses from adjacent unfrozen material.

At most of the involution localities described, the involuted materials are very inhomogeneous, and conditions are appropriate for differential frost heaving. When vertical ice veins melt, the resulting fissures might be expected to be filled, at least in part, from above. No such filled fissures have been found in involuted materials. Therefore, it seems likely that most of the ground-ice developed as layers approximately parallel to the bedding and that little indication of the location, size, and shape of any particular layer would remain after it had melted and the overlying material had slumped back. The forms of such subspheroidal involutions as those at locality 4 suggest interpenetration of masses of freezing, expanding material with adjacent plastic, unfrozen material. This is essentially the mode of development favored by Sharp (1942*a*, p. 130) for similar structures.

Role of perennially frozen ground.—Zeuner (1945, p. 12) states: "All varieties [of soils which owe their structures to intensive frost action] occur on permanently frozen sub-soil, and only certain types of stone-rings have been found in snow climates where tjaele [perennially frozen ground] is absent." Zeuner definitely includes involutions, which he calls "expansion trail," among the structures formed only on perennially frozen ground.

The involuted materials in Montana are at present well drained and have a relatively low water content. For intensive frost action to take place, it would be necessary to increase substantially the degree of saturation. The amount of precipitation in central Montana may have been greater during stages of glaciation than at present, but it is questionable whether this factor alone is sufficient. On the other hand, if the involutions formed in a zone of summer thaw overlying a perennially frozen layer, the precipitation and meltwater would not drain away readily. The presence of perennially frozen ground seems the most likely cause of a greater degree of saturation, since it would decrease the effectiveness of subsurface drainage.

Relation to stone polygons.—No surface phenomena, such as stone polygons, were observed to be associated with the involutions. If they had been present, their surface expression might well have been masked by the present soil zone. However, the character of the involutions does not indicate that any such phenomena were present in the past. The subsurface structures associated with stone polygons and similar features are roughly cellular and show segregation of coarse and fine particles (see references in n. 2). A "fossil" example is illustrated in Zeuner (1946, pl. XII-*A*). In contrast, the structures of the involutions are much less regular in form and were produced by the deformation of planes of stratification.

Relation to solifluction.—The term "solifluction" was proposed by Andersson (1906, pp. 95–96) for "the slow flowing from higher to lower ground of masses of waste saturated with water." He clearly recognized that solifluction, thus defined, was not restricted to areas of any one type of climate, but he maintained that a cold climate provided optimum conditions for its development. A recent description of solifluction in the Canadian arctic is given by Washburn (1947, pp. 88–96), who regards a slope of about 5° or more as necessary to bring about significant downslope movement. Most of the movement probably occurs when the material is very mobile as a result of spring melting of ground-ice in the seasonally thawed zone overlying perennially frozen ground. Some authors (such as Paterson, 1940, pp. 110–114) have extended the term "solifluction" to include processes which do not involve any downslope flow. Bryan (1946, p. 633) has proposed the term "congeliturbation" for all varieties of frost heaving and of solifluction induced by cold climate, thus including the formation of involutions in which no downslope flow has taken place.

The involutions at localities 2 and 4 and some of those at localities 1 and 3 were probably formed by local vertical movements caused by differential freezing and the formation of ground-ice. The presence of structures similar to drag folds (called "plications" by Bryan, 1946, p. 634) at localities 1 and 3 indicates that slight lateral movement occurred there, although the surface slopes are considerably less than 5° in inclination. At the time that the involutions were being formed, extensive solifluction (in the strict sense) must have been taking place on steeper slopes. The solifluction debris would become a structureless rubble, because of the mixing of surficial materials during movement. An example of such a rubble is found at the south end of the road cut at locality 8, on U.S. Highway 91, about 6.5 miles north of its junction with U.S. Highway 89 just east of Vaughn (highway distance). There the ground surface slopes at an angle of about 6°. Materials believed to be solifluction rubbles may be seen at a number of other localities.

Age of involutions.—The continental ice sheets of the two stages (or substages) of glaciation recorded in this area were of almost equal extent, and the climates of the two must have been quite similar. If formation of involutions occurred during one stage, it must have occurred during the other as well. As yet, no trustworthy evidence of more than one stage of formation of involutions has been discovered. This period of frost action may with reasonable certainty be correlated with the younger or Wisconsin ice advance, because of the thinness of the soil zone formed since the involutions developed. Furthermore, the involutions at locality 4 are developed in gravel interpreted by the writer as outwash of the Middle Wisconsin substage.

The ancient, now involuted, soil zone is similar in character to the modern soil zone and was formed in an interstadial or interglacial time previous to the period or periods of frost action responsible for the involutions. Since the ancient soil is several times as thick as the modern soil, it must have required a much longer period for its formation. These soils together constitute a complex, polygenetic soil zone, whose characteristics were "developed in more than one climatic regime" (Bryan and Albritton, 1943, p. 477).

Comparison with other areas.—In addition to the general description of involutions found in Zeuner's paper (1945, p. 8), numerous detailed accounts of European occurrences exist (see references in Sharp, 1942*a*). From gravels near Cambridge, England, Paterson (1940, pp. 110–121) has described globular involution-like structures, quite similar in form to those found at locality 4. The contents of the pockets commonly have a concentric arrangement, and the gravels in which the pockets occur may be festooned parallel to the walls of the pockets. Paterson believes that these structures are homologous with stone polygons and that the concentric arrangement is produced by ice-thrust from centers of freezing. However, the present writer agrees with Hollingworth (1940) that these structures more likely represent contortions of bedding unrelated to stone polygons and similar forms. They are probably similar in origin to the Montana involutions.

The first description of involutions in North America was by Denny (1936), who found contorted sands and disturbed gravels in southern Connecticut. Similar structures have been seen by the present writer at several localities in southern Rhode Island. Sharp (1942*a*) has described involutions from northeastern Illinois which are rather like the interpenetration of materials found at locality 4. Bryan and Ray (1940, p. 26) mention briefly an occurrence of involuted gravels in north-central Colorado.

DESCRIPTION OF WEDGE STRUCTURES

Localities 8 and 9 are at road cuts on U.S. Highway 91, about 6.5 and 5.6 miles, respectively, north of its junction with U.S. Highway 89 just east of Vaughn (highway distances). A detailed study was made of the structures on the east side of the highway at locality 8. Such structures also occur on the opposite side of the highway at locality 8 and at locality 9 and two near-by road cuts, one 0.4 mile north of locality 8 and the other 0.9 mile south of locality 9. Brief examination showed that the structures in these exposures were similar to those studied in detail.

The bedrock in this area is Colorado shale (Cretaceous), which is weathered to a very crumbly state to an average depth of 5–7 feet or more in the road cuts studied. At locality 8 it is mostly gray shale, with numerous thin orange-yellow

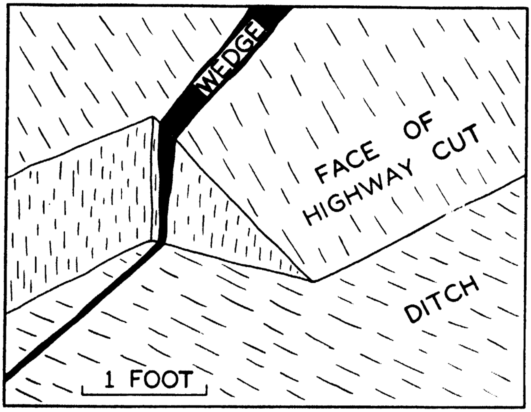

FIG. 5.—Generalized diagram of a wedge structure at locality 8. Wedge crops out as a band which descends the sloping face of the highway cut diagonally. That it is actually vertical is shown in a small excavation at the foot of the face.

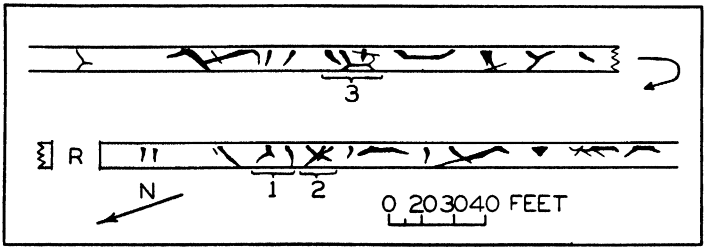

FIG. 6.—Sketch of plan of wedge structure exposed in the face of the cut on the east side of the highway at locality 8. Many small wedges are not shown. Symbol: *R*, break in exposure caused by an access road. Numbered groups of wedges are those illustrated in detail, as follows: *1*, pl. 3, *B*, and fig. 7; *2*, pl. 3, *C*, and fig. 8; *3*, fig. 9.

sandy layers. At locality 9 it consists largely of thin-bedded sandstone, much of which is silty. Both localities lie near the edge of a broad upland of very low relief which stands several hundred feet above a valley to the southwest. The structures are not restricted to this upland, however, since the cut 0.9 mile south of locality 9 lies near a valley bottom.

The material overlying bedrock is 2–3 feet thick and is probably mostly, if not entirely, derived from weathered bedrock. At locality 8 the mantle is largely sandy and contains many chips of thin-bedded sandstone, but in some parts of the exposure it consists mostly of shale debris. Scattered through the upper part of this mantle are pebbles of quartzite and limestone gravel, probably residual from the erosion of high-level gravel benches. Involutions are developed in the mantle at some places.

At some time in the past, fissures were developed in the upper, weathered part of the bedrock, and those fissures were subsequently filled with material from the mantle. These fissure fillings, which contain but a small quantity of shale fragments derived from the walls of the fissures, now appear on the cut faces as wedges, strongly contrasting in color and composition with the surrounding beds (pl. 3 and fig. 7). The wedges at these localities are essentially vertical in attitude, and the apparent dips of some of the wedges illustrated result from their exposure on a sloping rather than on a vertical surface. That part of the cut face represented in plate 3, *B* and *C*, and figures 7, 8, and 9 is 7–8 feet high; the lower part slopes at an angle of 30° or less, and the upper part rounds off to a nearly flat surface. Since these illustrations are views looking slightly downward at the exposures, they actually represent exposures on a sloping surface projected onto a plane inclined about 10° from the vertical. Thus wedges striking at right angles to the cut appear vertical; wedges striking parallel to the cut appear horizontal; and wedges striking obliquely to the cut appear to dip at various angles. An example of the last case is illustrated in figure 5.

Most of the wedges taper downward, as their name indicates, and some are conspicuously flared at their upper ends.

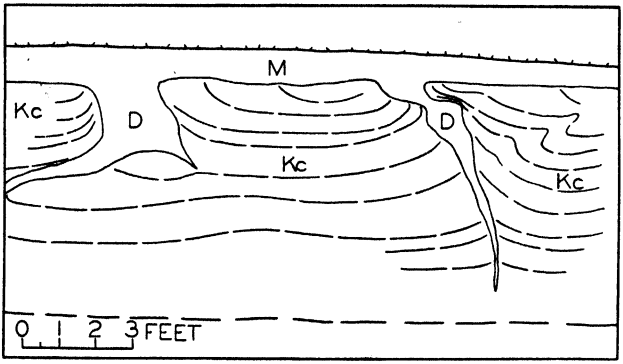

FIG. 7.—Diagram of the wedge structures at locality 8 shown in pl. 3, *B*. Symbols: *M*, surficial mantle of unconsolidated material; *D*, dark-colored surficial material filling fissures; *Kc*, Colorado shale, with bedding shown by broken lines. Heavy broken line in lower part of diagram represents base of face of highway cut.

Their walls may be straight or somewhat irregular. The majority extend no deeper than 6–8 feet below the surface. However, one large wedge at locality 9, shown in the right-hand part of plate 3, *A*, maintains a width of 1 foot to a depth of 13 feet below the surface and there ends abruptly just above a massive bed of sandstone. A few wedges appear to branch downward. The wedges vary in width from a fraction of an inch to about 3 feet. Some have a uniform width of an inch or less to a depth of several feet. Most of the wedges are relatively straight in plan, but some are gently curved. From a consideration of such groups of wedges as those shown in plate 3, *C*, and to the left of the large wedge in plate 3, *A*, it is apparent that they form an irregular polygonal network. Figure 6 shows a sketch of the plan of the wedges on the east side of the highway at locality 8.

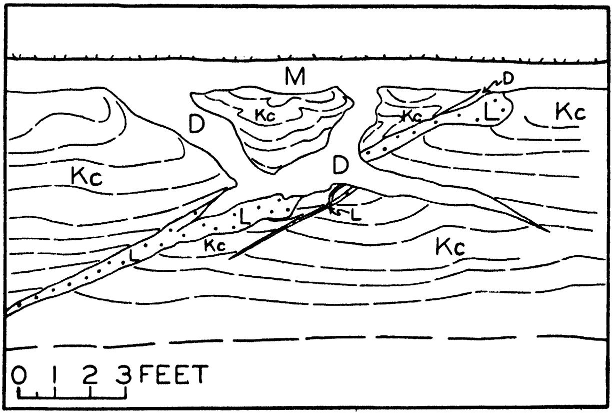

FIG. 8.—Diagram of the wedge structures at locality 8 shown in pl. 3, *C*. Symbols: *L*, light-colored surficial material (*dotted pattern*) filling older set of fissures, which are transected by younger set (*D*); other symbols as in fig. 7.

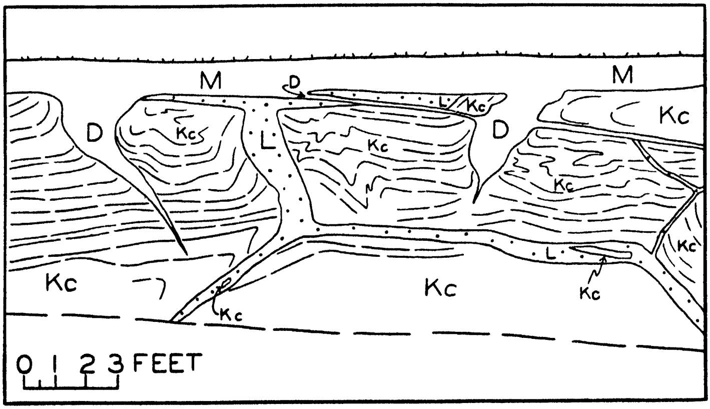

FIG. 9.—Diagram of wedge structures at locality 8, showing older set transected by younger set. Symbols as in fig. 8.

This sketch is not accurate in detail but represents the general relations of the wedges. The average diameter of the polygons is at least 25 feet and probably more. Since the horizontal breadth of exposure on the cut face is only 15 feet and since most of the wedges do not extend deeply enough to be exposed to the foot of the face, not many junctions of wedges appear in the sketch. The narrow lower ends of a few of the deeper wedges can be traced across the floor of the ditch to the edge of the highway grade, a distance of 12 feet. No relation between the orientation of the wedges and that of the joint systems in the shale is apparent, because the jointing of the shale has been obscured by the weathering and deformation which it has undergone.

The normally horizontal bedding of the shale is in all cases turned upward along wedges and in some places is intensely deformed. Deformation is generally less along the narrower, lower parts of wedges but is found along narrow wedges as well as broad ones. This strongly suggests that the fissures were widened by pressure exerted from within. It seems unlikely that the present contents of the fissures could exert this pressure.

Almost all junctions of wedges show no transecting relationships, and the materials of the joining wedges are similar in color and in grain size. Most of the wedges consist of dark-brown, relatively fine-grained material, derived from the carbonaceous surface zone (pl. 3, *B*, and fig. 7); but some of the narrow, deep-seated wedges are composed of light-colored sand. In a few groups of wedges there also exist large wedges which are filled principally with light-colored sandy material and which are sharply transected by the darker wedges (figs. 8 and 9). Although the lower part of the mantle is lighter in color and sandier than the upper part, it is not so light as is the material filling the light-colored wedges, which are practically invisible in photographs (pl. 3, *C*). Furthermore, the material of the light-colored wedges seems to be separated by a relatively sharp contact from the lower part of the mantle. If these wedges are considerably older than the darker ones, it may be that an original dark filling has been "bleached" of its organic matter in the intervening period. It may be significant that both the two light-colored wedges illustrated and the one other example found extend several feet deeper than do the darker wedges, which cut across them.

A situation somewhat different from that at localities 8 and 9 is found at locality 10, a road cut on U.S. Highway 89, about 3.5 miles southeast of Choteau (highway distance), in the center of the northeast quarter of Section 8, T. 23 N., R. 4. W. The bedrock is black shale of the Colorado formation. It is overlain by sandy slope-wash material derived from an outlier, 1 mile to the southwest, of an escarpment of Virgelle (Eagle) sandstone (Montana group, Cretaceous), which overlies the Colorado shale. Part of the exposure in the cut on the west side of the highway is shown in figure 10. Since the cut face slopes at an angle of about 70°, the diagram represents a nearly vertical section. Cutting back of the face confirms the fact that the bands of sand which appear horizontal in section are actually horizontal, unlike any of those observed at localities 8 and 9. Therefore, not only vertical but also horizontal fissures were formed in the weathered shale at this locality. No cross-cutting relationships between the vertical wedges and horizontal layers are present, and all the fissures must have developed as a single set and later been filled together.

The bedding in much of the shale cannot be readily traced, but the degree of deformation is shown by the disruption which an orange-yellow bentonitic layer has undergone. Some of the shale masses which appear to be isolated from the main body of shale in figure 10 are perhaps connected with it in a direction at right angles to the plane of the section. Similar associations of vertical wedges with horizontal layers are shown in road cuts 0.3 mile north and south along the highway.

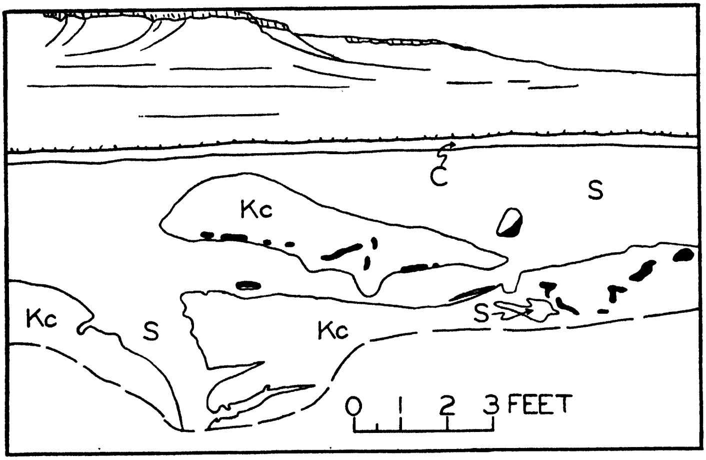

FIG. 10.—Diagram of wedge structures at locality 10, showing horizontal fissures as well as vertical. Above crest of nearly vertical exposure is a smooth, gently sloping surface, extending 1 mile to the southwest to a sandstone escarpment. Symbols: *C*, the humus-rich A horizon of the modern soil; *S*, sandy slope-wash material; *Kc*, Colorado shale. Small bodies shown in solid black are remnants of a disturbed bentonitic layer in the shale.

No other occurrences of wedge structures, with or without horizontal layers, have been seen by the writer within the area studied.

ORIGIN OF WEDGE STRUCTURES

COMPARISON WITH ICE-VEIN STRUCTURES

The most extensive description available of the ice veins which exist in arctic and subarctic regions is that by Taber (1943, pp. 1510–1529). Vertical ice veins and horizontal ice layers or lenses constitute the main modes of occurrence of segregated ground-ice in perennially frozen ground in the Alaskan areas which he has studied. Most veins have simple wedge-shaped sections (some branching downward), but some flare at their upper ends or are confluent above with ice layers. Taber believes that all ice veins, when first formed, extended downward from ice layers and that, where the layers are absent, they have been removed by later thawing and erosion. Rarely are isolated veins found. Most of the veins occur in polygonal networks, and the average diameters of polygons range from 20 to 45 feet in different areas. Where networks of ice veins join ice layers below as well as above, a more or less cellular structure is produced. The bedding of the material adjacent to the veins is commonly distorted by pressure exerted by the vein during its growth.

This description of Alaskan ice veins matches that of the Montana wedge structures very closely, except, of course, for the fact that any ice veins once present in Montana have thawed and the fissures which they occupied have been filled by material which slumped in from the surface. Taber notes the slumping of surface material into the upper parts of fissures occupied by ice veins which are gradually thawing downward (Taber, 1943, pl. 13, fig. 1; pl. 15, figs. 1 and 2; pl. 16, figs. 1 and 2). This is a strong argument for the essential identity of the two sets of structures, especially since the slump-filled portions of some of the fissures extend to a depth several times greater than that reached by any of the Montana wedge structures. Alaskan ice veins attain greater width and depth than do the Montana wedges, but this is to be expected of an area of cooler climate.

The horizontal filled fissures found at locality 10 indicate that horizontal layers of ice were formed there in association with the vertical veins. There is little evidence bearing on Taber's suggestion that all ice veins form below ice layers, but it may well be that the ice layer which presumably lay beneath the largest isolated mass of shale shown in figure 10 extended some distance laterally at or near the base of the mantle.

Wedge structures attributed to the filling of fissures once occupied by ice veins have been found at a number of localities in Europe. Zeuner (1945, pp. 10–12) refers very briefly to some examples. Paterson (1940, pp. 102–105) describes from near Cambridge, England, some wedge structures which are, in plan, independent linear features which cut across one another at various angles, unlike those described above. The wedge structures in central Germany described as "pseudomorphs after Pleistocene ice wedges" by Selzer (1936, p. 286) are quite similar to those in Montana. The irregular network which they form (Selzer, 1936, pp. 282–283) is more closely similar to that of the Montana wedges than is the relatively regular polygonal network which Leffingwell (1919, p. 210) found in the case of existing ice veins in northern Alaska.

All the authors cited in the last paragraph subscribe to the frost-crack hypothesis of formation of ice veins (Leffingwell, 1919, p. 206). Taber, however, presents a very convincing argument for the concurrent formation of ice layers and ice veins by segregation of ground-ice (1943, pp. 1519–1528). Each hypothesis requires the presence of perennially frozen ground. The Montana wedge structures offer no evidence as to which hypothesis is more nearly correct.

ALTERNATIVE MODES OF ORIGIN

Several of the hypotheses of origin of involutions which were considered and rejected may be summarily rejected as possible modes of origin of wedge structures. These include volume change in bentonite, differential loading, and tectonic deformation.

None of the wedge structure localities lies within a glaciated area, so that overriding by glacial ice need not be considered. However, at all the localities at which wedge structures were seen, ice-rafted pebbles or cobbles of crystalline rocks of Canadian origin occur on the ground surface. Therefore, the possibility of the production of the structures by the grounding of the icebergs of a proglacial lake in shallow water must be considered. It is unlikely, however, that dragging ice masses could cause the formation of cracks extending to a maximum depth of at least 13 feet below the ground surface. It seems impossible that such an agent could produce the deformation of bedding planes shown near the walls of the fissures, strongly suggesting the exertion of pressure from within, or that it could cause the arrangement of the fissures in a polygonal network.

Masses of rock near or at the ground surface may, under certain conditions, tilt in a downslope direction, producing wedge-shaped cracks widening upward, which tend to be filled with surficial material. Several varieties of this type of movement have been discussed by Judson (1947). The following relations show that the Montana wedge structures cannot have been formed in this way. Mass movement of the type described, which seems to have been assisted by ice-wedging at some places, occurs only in blocks of a competent bed; but the weathered shale of the Colorado formation is hardly capable of acting as a coherent block. Cracks that form by such movement tend to extend parallel to the surface contours, whereas the present fissures occur in an irregular network. The Colorado shale shows no tilting parallel to the direction of slope at the wedge structure localities, and in any case the structures have been found only where the surface slopes at an angle of less than 5°.

Lupher (1944) has described clastic dikes in Pleistocene sediments of the Columbia Basin region. These dikes were, like the Montana wedge structures, formed under periglacial conditions, but there is no evidence that they are the results of thawing of ice wedges (Lupher, 1944, p. 1458). The dikes, many of which are multiple, are parallel-walled rather than wedge-shaped and extend to a maximum depth of 120 feet. The adjacent materials are undeformed. The fissures occupied by the dikes are believed to have formed principally because of subsidence as buried ice masses and frozen sediments thawed, and they were filled from above.

AGE OF WEDGE STRUCTURES

Wedge structures of different ages are described by Selzer (1936, pp. 279–280). They were formed in a loess concurrently with its deposition and at successively higher levels as the loess accumulated. No such situation existed at locality 8. The two sets of wedges there differ markedly in the lithology of the material filling the fissures, and the systematic difference in size may be important also. The writer believes that a substantial period of time elapsed between the formation of the two sets. Since the cooling of climate necessary for the development of ice veins presumably occurred at a time of extension of glaciers, there is a strong temptation to assume a correlation between the two stages of wedge formation and the two stages (or substages) of glaciation of which evidence is recorded in the area. This correlation is plausible, but there is no direct evidence to support it.

THE PERIGLACIAL CLIMATE

The use of the term "periglacial" has been criticized by Flint (1947, p. 461) as follows:

> The association of flow earth [solifluction debris], loess, and ventifacts has been considered as indicating "periglacial" conditions, by which are implied very cold, windy conditions in the neighborhood of an ice-sheet margin. This general term can have no absolute quantitative significance; strictly speaking, a periglacial climate is any climate that exists around the edge of a large glacier and might be of more than one kind. Progressive change of periglacial climate has been specifically recognized....

This is a valid criticism; for, if uncritical use is made of the term, it becomes almost meaningless. The climate of the area immediately surrounding an ice sheet varies not only in space but also in time. Considerable evidence has been found which indicates that, as the last ice sheets in Europe and North America waned, their influence over the climate of the areas immediately surrounding them waned also. For example, Ahlmann (1947, p. 129) says: "At the time when the inland ice melted away from South Sweden the high polar climate had already finished. This is, among other things, made probable by the absence of peri-glacial solifluction and stone polygons on the Scandinavian Peninsula." In contrast, periglacial features of these and other kinds are common in many areas of central and western Europe. The climate implied by the term "periglacial" has been extended into temperate and cool-temperate regions in the past only as an attendant circumstance of glaciation by large ice sheets, and the periglacial climate may have come to an end in those regions before the ice sheets had completely withdrawn. Therefore, it must be understood that the term "periglacial" as it is used at present refers to a specific set of conditions believed to have been characteristic of a relatively narrow zone at times of maximum or near-maximum glaciation and extending farther away from large glaciers on highlands than in lower areas. A term with this significance is undeniably convenient, and, if this one is to be abandoned, it should be abandoned only in favor of one more specific.

It has been argued in the section on the origin of involutions that the presence of involutions definitely suggests the former existence of perennially frozen ground or tjaele in central Montana. The presence of ice-vein structures may be taken as proof of this point, since ice veins develop only in tjaele. The wide distribution of involutions and wedge structures in the area, in a variety of topographic and geologic situations, shows that the conditions promoting their development were general and not limited to a few favorable localities. If tjaele developed at localities 8 and 9, with practically maximum exposure to insolation, it must have been essentially continuous throughout the area.

Tjaele develops only where the mean annual temperature is below 0° C., perhaps several degrees below (Taber, 1943, p. 1506; Muller, 1947, p. 4). The present mean annual temperatures of several localities in central Montana range from 5° to 8° C. This indicates a temperature change of 8° C. or more since the time that the wedge structures were formed. The present southern boundary of tjaele east of the Rocky Mountains is between 57° and 60° N. Lat. (Muller, 1947, p. 5), or about 250–400 miles north of the latitude of central Montana.

There is no obvious way of estimating the depth to which tjaele formerly existed in central Montana. The development of broad ice veins is probably limited downward by the greater strength of unweathered rock. Ice veins do develop in solid rock (Taber, 1943, pp. 1528–1529); in fact, they extend deeper in rock than in unconsolidated material, because the greater conductivity of rock results in a greater depth of freezing. However, these veins are generally relatively narrow, and it is doubtful whether any recognizable evidence of their former existence would remain after they had thawed.

The upper ends of the wedge structures represent the upper limit of tjaele and therefore the lower limit of the seasonally thawed or "active" layer. At localities 8 and 9 the active layer seems to have extended to a depth of about 3 feet, and the ground was perennially frozen to a depth of at least 13 feet and almost certainly substantially deeper. Involutions, on the other hand, are formed not in tjaele but in the active layer. Since involutions at a number of localities extend to a depth of about 8 feet, the active layer there must have been at least that thick. These local differences in the depth of thaw may have been caused by differences in type of material or vegetative cover or any of several other factors (Muller, 1947, pp. 6–10). The average annual depth of frost penetration in central Montana at present is about 3 feet.

Wind action has been considered by many authors to be frequently associated with intensive frost action in the periglacial zone (Flint, 1947, p. 461). Among the authors cited elsewhere in this paper, Denny (1936) describes the association of fine-grained wind-blown material containing ventifacts with involutions in southern Connecticut; and Selzer (1936) discusses ice-wedge structures that were formed concurrently with the deposition of loess. Wind-polished stones are present on the ground surface at many places in central Montana and occur at localities 1 and 8 and very near locality 10. The wind action responsible for the polishing of these stones took place at some time in the past when there was a less nearly continuous cover of grass, since the vegetation effectively inhibits wind action at the present time. It may have occurred under periglacial conditions, or it may have taken place at a time of drier climate than that of the present, subsequent to the retreat of the last ice sheet.

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PLEISTOCENE VENTIFACTS EAST OF THE BIG HORN MOUNTAINS, WYOMING[1]

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ABSTRACT

Ancient ventifacts are abundant on gravel-mantled erosion surfaces along the eastern base of the Big Horn Mountains. Stones of varied lithologic characteristics and from less than 1 inch to 6 feet in diameter have wind-cut surfaces displaying the pitting, fluting, grooving, and luster that is characteristic of ventifacts. Faces making angles greater than 55° with the wind are pitted, and faces at lower angles are grooved and fluted. As many as twenty separate faces have been cut on a single stone. In this area such multiple faces cannot be explained by wind-splitting or by variable winds, so they are attributed to changes in stone position caused largely by congeliturbation (frost action) and wind scour.

Ventifacts rest on erosion surfaces at four levels, 25–325 feet above stream grade. On the lower surfaces wind-cut stones are abundant and fresh, but on the higher surfaces they are sparse and deeply weathered. At least four separate periods of wind-cutting are indicated by this evidence. Cutting is attributed to sand picked up from barren flats and hurled against stones lying in the same environment. Conditions most favorable for wind-cutting prevailed during times of glaciation in the mountains, when winds were strong and floods of meltwater inundated parts of the piedmont, producing broad barren flats mantled by sand and gravel. Ventifacts on the lower erosion surfaces are probably Wisconsin, and those on the highest surfaces may be pre-Wisconsin.

Pleistocene wind directions have been measured at twenty-nine separate localities by careful reference to wind-cut faces on large, presumably stable, boulders. The mean Pleistocene wind direction so determined is N. 29° W. Modern winds are also consistently from the northwest, and it appears that local orographic control was supreme in the Pleistocene as now. Neither glaciers in the Big Horn Mountains nor the continental ice sheet, 250 miles north, exerted much influence on local wind directions.

INTRODUCTION

INTRODUCTORY STATEMENT

This article describes ancient ventifacts east of the Big Horn Mountains, Wyoming, analyzes their mode of origin, considers the conditions under which they formed, gives an approximate date for the wind-cutting, and demonstrates the use of ventifacts in determining ancient wind directions. Ventifacts have not previously been reported from this area; Gale and Wegemann (1910, p. 138) mistook them for glaciated stones.

Study of erosion surfaces east of the Big Horns was started in 1940, and special attention was devoted to ventifacts in 1946. Total field time to date is 7 weeks. The area is covered by the Fort McKinney, Sheridan, and Dayton topographic quadrangles and by aerial photographs from the United States Department of Agriculture.

LOCATION AND PHYSICAL SETTING

The Big Horn Mountains are the principal front range of the Middle Rockies in north-central Wyoming. The area studied lies along the east base of this range between Buffalo and Dayton (fig. 1).

A succession of erosion surfaces (pl. 1, *A*) flanks the east base of the range and extends at least 15 miles eastward into Powder River Basin. They have altitudes of 4,200–6,000 feet and are 25–375 feet above stream grade. The higher, more extensive members are similar to features called "pediments" farther south (Bryan and Ray, 1940, p. 20), and the lower surfaces are terraces bordering the present streams. Since study of these surfaces is not yet completed, a noncom-

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mittal terminology is preferred, and they will be referred to simply as "piedmont erosion surfaces." Alden (1932, pp. 55–56) classified them into two groups corresponding to his No. 2 and No. 3 benches. His No. 1 bench, or Flaxville Plain, may also be represented locally by a few small remnants. Detailed mapping and numerous measurements by hand level show that Alden's classification, although oversimplified, can be used with the addition of plus and minus signs to indicate the relationship of surfaces somewhat above and below the best-developed levels of the No. 2 and No. 3 benches.

These erosion surfaces truncate steeply tilted Mesozoic strata (pl. 1, *A*) and more nearly horizontal Tertiary beds; only the lowest terrace appears to be an alluvial fill. The surfaces cut on bedrock are mantled by boulderly stream gravels 5–10 feet thick, and it is the surficial stones of this mantle that are wind-cut. Ventifacts have been reported in similar settings from other piedmont areas in the Rocky Mountains (Stone, 1889, p. 421; Mehl, 1925; Delo, 1930; Wentworth and Dickey, 1935, p. 100; Schoewe, 1938; Bryan and Ray, 1940, pp. 22, 26).

DISTRIBUTION OF VENTIFACTS

Ventifacts are most abundant on erosion surfaces immediately north of Buffalo (fig. 1), but specimens are scattered for 65 miles along the east front of the Big Horn Mountains from Trabing to Ranchester (fig. 1) and probably beyond in areas not investigated. The wind arrows plotted on figure 1 represent only a fraction of the total localities; for wind directions could not be determined reliably at all places. Wind-cut stones were found in a broad belt about 12 miles wide, east of the mountains, but they are most abundant in the central part of this belt, 4–7 miles from the base of the mountains. Only old ventifacts on high surfaces are close to the mountain front; young ventifacts on low surfaces are usually several miles farther east. A few wind-cut stones were found on the subsummit surface (fig. 6) south of Sister's Hill at 7,700 feet altitude. Ray (1940, p. 1857) reports ventifacts in a similar upland situation in Colorado.

Ventifacts are lacking in parts of the piedmont. Their absence north of Tongue River is probably due to the fact that the rocks composing gravels on these erosion surfaces are not resistant enough to preserve evidence of ancient wind-cutting. Erosion surfaces in the vicinity of Sheridan are buried beneath fine sheet wash or flood-plain silts, which effectively hide the ventifacts, if any, except possibly in cut banks where none has yet been found. Many of the highest erosion surfaces are without ventifacts because of the scarcity of dense siliceous stones which preserve evidence of wind-cutting in spite of extended weathering. Finally, ventifacts were not found in many places presumably because the stones were protected from the wind by vegetation or topography.

Ventifacts lie on erosion surfaces 25–325 feet above stream grade. In places acres of ground are littered with wind-cut stones.[2] Most of them are partly buried in fine soily material, probably a sheet-wash deposit. Only near the base of

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[2] Anyone wishing to collect or study ventifacts will find them especially abundant at the following easily accessible localities.

1. Follow U.S. Highway 87 for 2.1 miles north from the courthouse in Buffalo. At the crest of the rise, before the highway descends to Rock Creek (Fort McKinney quadrangle), a dirt road branches right to an abandoned gravel pit, 100 yards east. Ventifacts are scattered over the surface south of the gravel pit (SE. ¼ of the NE. ¼, Sec. 22, T. 51 N., R. 82 W.).

2. This locality lies on a low terrace just north

steep slopes is the wash thick enough to bury the ventifacts completely. Examination of a number of cuts and excavations showed that only the surficial stones of the gravel veneer are wind-cut.

FEATURES OF THE VENTIFACTS

GENERAL STATEMENT

In the following discussion the reader should keep two facts in mind. First, Big Horn ventifacts were cut by wind blowing consistently from one direction, the northwest. This is demonstrated by numerous large boulders having but a single wind-cut surface, which, in every instance, faces northwest. These boulders are too large to have been moved appreciably by processes which repeatedly shifted smaller stones. Second, wind-cut stones 6–15 inches in diameter are abundant in this region, and much of the following material applies to ventifacts of this size or larger. This distinction is significant because many previous generalizations concerning ventifacts have been made with respect to smaller stones.

ROCK TYPES

Big Horn ventifacts are composed of chert, quartz, quartzite, quartzitic sandstone, gneiss, hornfels, pegmatite, diabase, and many varieties of granite. No evidence of wind-cutting was found on limestone or dolomite, although ventifacts of this composition are common elsewhere (Wade, 1910, p. 395; Hume, 1921, p. 252; King, 1936*a*, p. 203; Maxson, 1940, pp. 726–727). The efficacy of solution weathering (Bryan, 1929; Scott, 1947) on carbonate rocks in this environment is demonstrated by abundant solution-faceted stones (pl. 4, *B*), and the lack of limestone and dolomite ventifacts can be attributed to destruction of wind-cut surfaces by solution.

The ventifacts are not all of the same age. The oldest are limited largely to dense siliceous rocks, resistant to weathering, upon which only small patches of the wind-cut faces are preserved (pl. 1, *B*). The youngest are composed of all the rocks listed above, with granitic types predominating. Details of wind-cut surfaces are influenced by lithologic characteristics and structure, as subsequently noted.

SIZE AND SHAPE

Most Big Horn ventifacts are from a few inches to a foot or two in diameter, and extremes range from pebbles $\frac{7}{8}$ inch long to boulders 6 feet in diameter. Somewhat smaller (King, 1936*b*, p. 277; Thiesmeyer and Digman, 1942, p. 179) and considerably larger (Powers, 1936, p. 215; Mather, Goldthwait, and Thiesmeyer, 1942, p. 1166) wind-cut stones are reported elsewhere.

Big Horn ventifacts have up to twenty wind-cut faces, and the shapes include most of those previously described (Bryan, 1931, p. 30; Thiesmeyer and Digman, 1942, p. 180). Boulders projecting several feet above the ground usually show a curved, wind-carved surface (pl. 4, *A*), but a few exhibit a more nearly plane, wind-cut face (pl. 1, *C*). Ventifacts of intermediate size, 6–15 inches in diameter, are mostly ridge-shaped, owing to the intersection of two major opposed faces, or pyramidal, with three to six faces converging toward an apex (pl. 3, *A*). Wind-cut faces on most pyramidal and ridge-shaped ventifacts are confined to the upper part of the stone, probably because the stone was partly embedded

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of the Buffalo fair grounds. Drive north 1.7 miles on U.S. Highway 87 from the courthouse in Buffalo and turn east on the dirt road where the highway first rises onto the terrace. Ventifacts are scattered over the surface for at least 1 mile eastward along this road (SW. corner, Sec. 23, T. 51 N., R. 82 W.).

in the ground when cutting occurred. The ventifacts with as many as ten to twenty faces are 8–12 inches in diameter and exhibit cutting practically all over (pl. 3, *C*). Their shapes can best be described as irregular or polygonal, and the edges between many faces are in some instances arranged approximately in a zone like the edges of a mineral crystal (King, 1936*a*, p. 208). Smaller stones with cutting on all sides have fewer faces.

PITS, FLUTES, AND GROOVES

In detail, nearly all wind-cut faces show some degree of pitting, fluting, or grooving (pl. 3). The nature of these markings and their degree of development depends primarily upon lithologic characteristics and orientation of the face to the wind direction. Pits are irregularly shaped (pl. 3, *E*) and from the size of a pencil point to an inch in diameter and depth. In the Big Horn region they are best developed in medium- to coarse-grained crystalline rocks, although elsewhere common in lava (Gilbert, 1875, p. 84; Blackwelder, 1929*b*, p. 256; Denny, 1941, p. 257). Pits are strongly influenced by mineral resistance and rock structure. Each projecting point on a pitted granitic surface is capped by an area of quartz or other hard mineral, as long ago noted by Blake (1855, p. 179) and Brown (1909, pp. 130–131). Pits form on faces making angles between 55° and 90° with the wind. On curved faces, a transition from pits to deep flutes with overhanging ends occurs as the angle between the face and wind decreases (pl. 4, *A*).

Flutes are scoop-shaped in plan and broadly U-shaped in cross section. They may be so small as to be nearly indiscernible to the naked eye or up to 6 inches long, 1½ inches wide, and ¾ inch

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PLATE 1

A, Piedmont erosion surfaces, south from Bald Ridge. Chiefly gravel-mantled No. 2 surface truncating tilted Mesozoic beds.

B, Extensively weathered ventifact on No. 2 surface. Only small patches of wind-cut surface remain on this quartzite boulder.

C, Granite boulder with wind-cut face. On terrace of No. 3 group north of fair grounds at Buffalo.

PLATE 2

A, Wind-carved boulders on terrace of No. 3 group north of Buffalo fair grounds. Wind blew from the lower right (northwest).

B, Wind-cut grooves in gneissic granite. Foliation parallels northwest wind, which blew from lower left to upper right. Six-inch ruler on boulder.

PLATE 3

A, Three-facet pyramidal ventifact on granite. Pocket knife in these photographs is 3⅛ inches long.

B, Fluted wind-cut face on gneiss with fluting transverse to foliation.

C, Multiple-face ventifact on granite. End view showing cutting on all sides in nearly vertical zone.

D, Top view of polygonal ventifact in granite. Upper grooved face curves downward to the right.

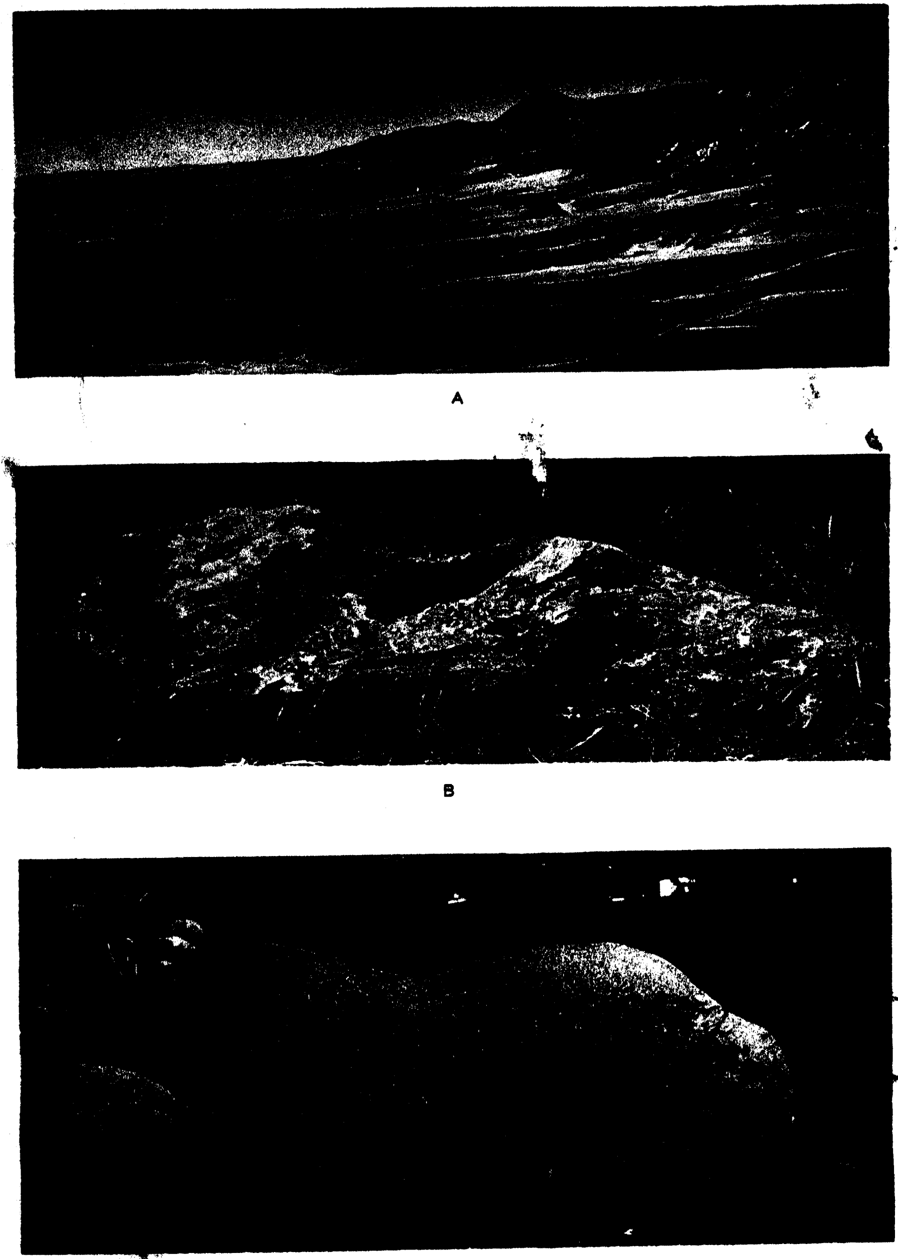
E, Pitted and grooved faces on pegmatite.

F, Wind-cut flutes on quartzitic sandstone, nearly as large but not so deep as stream flutes.

PLATE 4

A, Pits and flutes on granitic boulder, showing curved wind-cut face with radial arrangement of flutes.

B, Solution-faceted limestone cobble. White material on under side is secondary calcareous deposit. Pocket knife 3⅛ inches long.



A

B

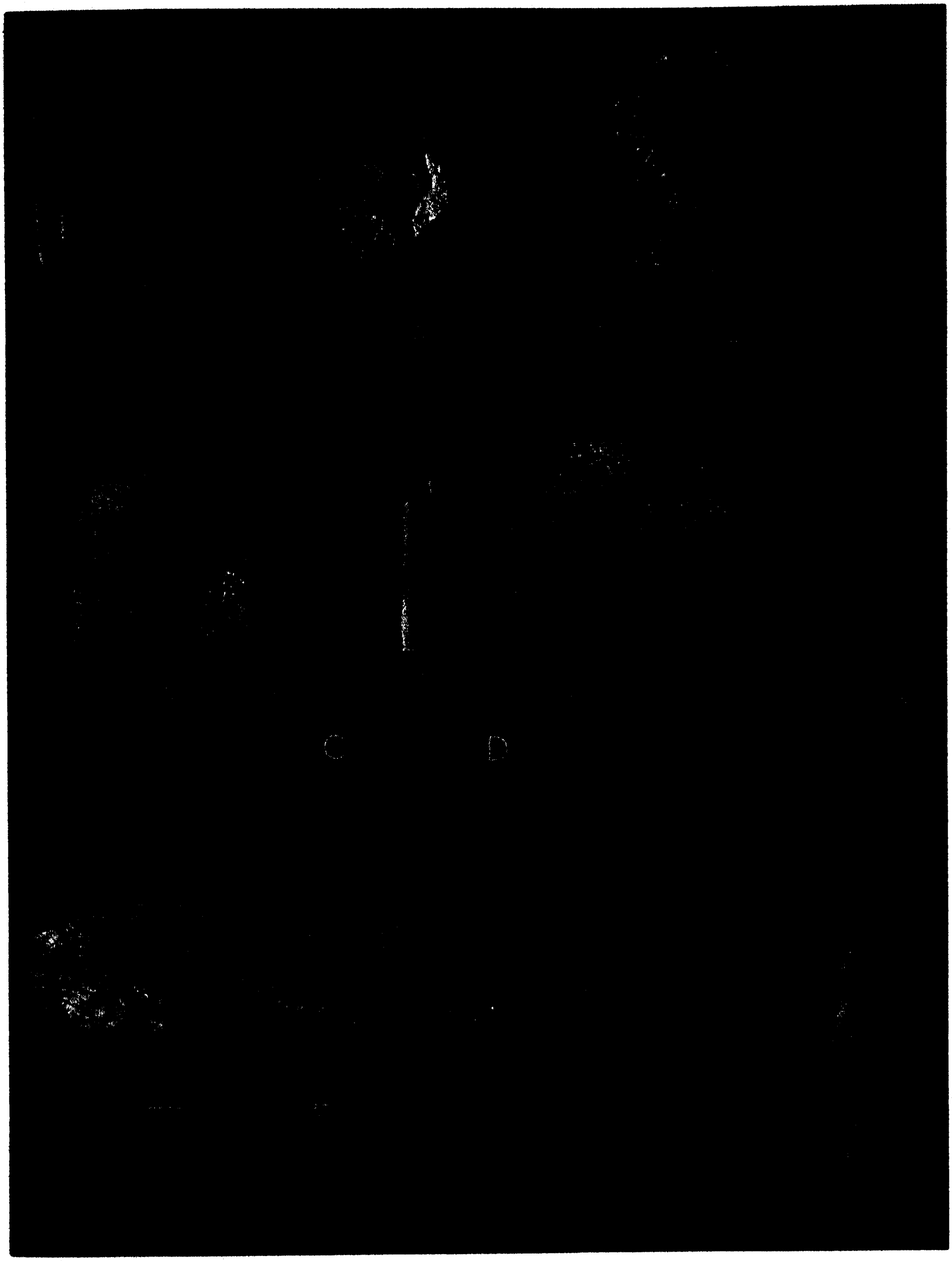
C

Piedmont erosion surface, wind-cut and weathered boulders

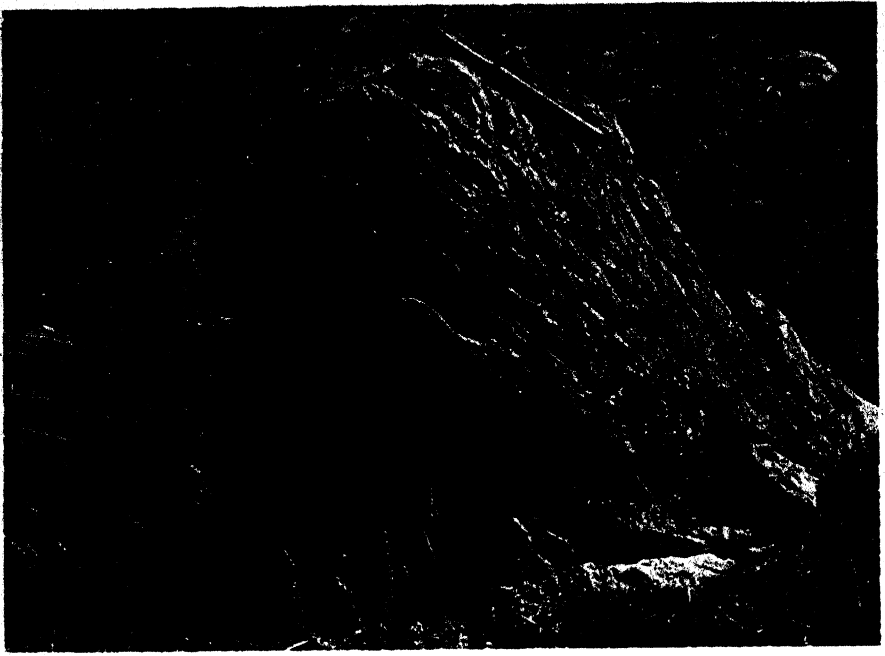
A

B

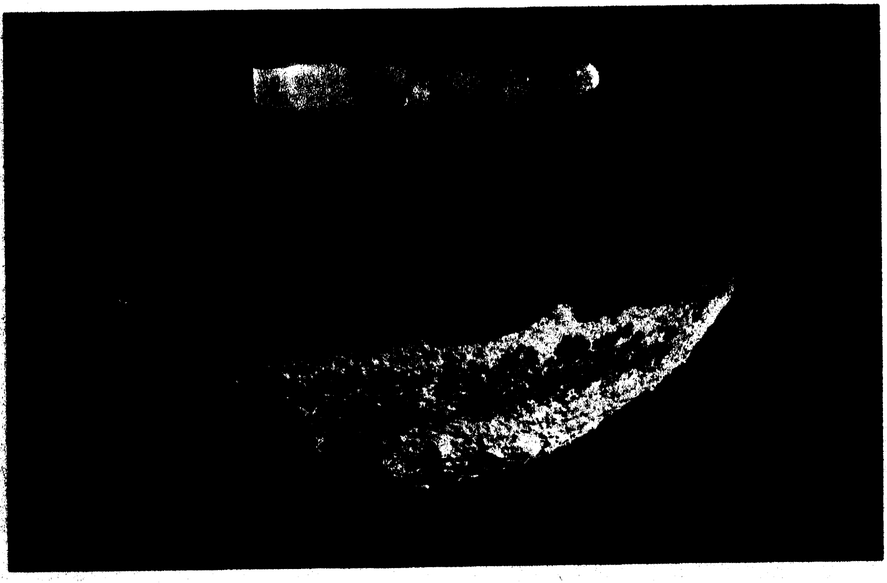
Wind-carved boulders



Pleistocene ventifacts



A



B

Pitted and fluted granite boulder, and solution-faceted limestone cobble deep. The exact shape of a flute is determined largely by intersection with its neighbors; but it is usually open at one end and closed at the other. Most flutes open downwind, but some with exceptional depth have overhanging sides and a pit with an overhanging roof at the downwind end. Flutes form best on surfaces inclined less than 40° to the wind, although fluted surfaces inclined up to 55° have been seen. Flutes become shorter and deeper as the inclination of the surface steepens. Unlike pits, they are strikingly independent of mineral hardness and rock structure (pl. 3, *B*) unless structure happens to parallel wind direction.

Grooves are longer than flutes and open at both ends. The largest seen were 18 inches long, 2 inches wide, and 1 inch deep (pl. 2, *B*). They are best developed on surfaces gently inclined or parallel to the wind and, like flutes, may cut indiscriminately across mineral grains and rock structures (pl. 3, *E*). The change from flutes to grooves is gradational, and the two forms are not mutually exclusive, for the surfaces of large grooves are often fluted on a small scale.

Maxson (1940, pp. 727, 733) attributes flutes to cutting by wind vortices and vortex trains, but this requires cutting particles fine enough to follow vortex currents. Most sand grains traveling by saltation probably pass through small vortices without much change of path. However, the polish on many wind-cut surfaces suggests that the wind carries some material finer than most wind-blown sand; and Cailleux (1942, p. 54) thinks that fine material may help cut ventifacts. This material may be fine enough to follow wind currents in vortices and thereby aid in cutting flutes and grooves. Schoewe (1932, p. 127) discovered that sand grains impinging at low angles on hard, smooth surfaces skid instead of rebounding directly into the air. This could occur on wind-cut rock surfaces and may also have something to do with the development of flutes and grooves.

The flutes or grooves on a face are parallel except on large curved surfaces facing into the wind, where they radiate outward from a central pitted area (pl. 4, *A*). Markings on adjacent faces are usually not parallel and may be perpendicular, parallel, or oblique to the edges between the faces. This discordance can be ascribed largely to changes in position of the stone between periods of cutting or, in limited degree, to the different orientation of faces subjected to contemporaneous cutting. Flutes cut on inclined faces at right angles to wind direction are parallel to the wind, but this is not true of markings cut on inclined faces oblique to the wind; for the direction of a wind current is partly controlled by the orientation of the face across which it blows.

Lithologic characteristics influence the nature of a wind-cut face and its markings. Dense resistant rocks, such as quartz and chert, have especially smooth faces, with shallow grooves, flutes, and pits. Faces on fine-grained diabase and similar rocks are more conspicuously marked than quartz, but their markings do not begin to compare with those on granitic and gneissic rocks. Coarse gneisses have exceptionally deep grooves and pits, where foliation is parallel to the wind (pl. 2, *B*). Faces on a particular quartzitic sandstone of this region have especially broad, shallow flutes (pl. 3, *F*), which, contrary to Maxson's statement (1940, p. 727), are as large, although not so deep, as some stream flutes.

WIND-CUT SURFACES

Wind-cut surfaces are distinguished from surfaces of other origin primarily by small-scale pitting, fluting, or grooving and a degree of luster different from that normally displayed by the stone (Cailleux, 1942, p. 58; Thiesmeyer, 1942, p. 244). Among students of ventifacts the term "facet" appears to mean a relatively plane surface cut at right angles to the wind, regardless of the original shape of the stone. Since many wind-cut surfaces on boulders in this area lack the planar qualities of a facet or were probably formed by modification of pre-existing surfaces not at right angles to the wind, they are described as "faces." A facet is simply one particular type of face.

As previously shown (Gilbert, 1875, p. 84; Blackwelder, 1929*a*, p. 396; 1929*b*, p. 256; Denny, 1941, p. 257; and many others), large boulders projecting several feet above the ground may split the wind in radial fashion, so that a curved surface is formed. In the background of plate 2, *A*, is such a boulder, but in the foreground is another large stone with a more gently inclined, nearly plane wind-cut face—in fact, a facet. It is conceivable that the curved surface is the initial stage and the facet the advanced stage of long-continued wind-cutting on large stones, but the various transition stages between these extremes were not observed. Since, in this instance, the two boulders are lithologically similar and must have been exposed to wind-cutting for about the same length of time, it would seem that original shape has been the controlling factor. The faceted boulder probably had a surface about parallel to the present facet before wind-cutting occurred, or possibly the boulder was cut more rapidly because it was low and close to the ground where sandblasting is most intense.

Some wind-cut faces on stones of intermediate size, 6–15 inches in diameter, have all the aspects of true facets; but many others are gently concave (fig. 2), convex, or even wavy, and some curve through angles of 70°–80° (pl. 3, *D*). Curves in sections transverse to the wind are predominantly convex, but in sections parallel to the wind they may be either concave or convex. Wind-cut surfaces on some ventifacts appear curved because they consist of a series of small facets intersecting in very oblique angles.

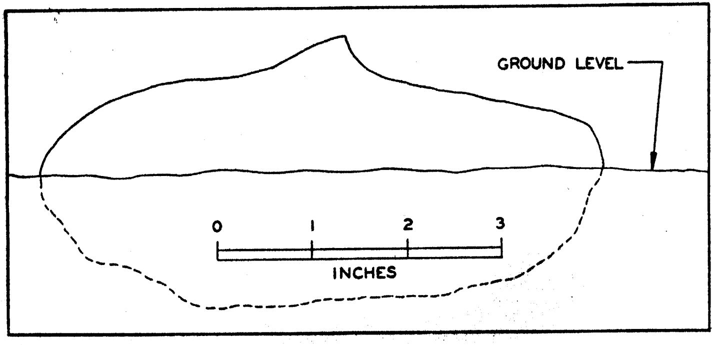


FIG. 2.—Concave faces on quartz ventifact

Bryan (1931, p. 35) and Schoewe (1932, pp. 131–132) have emphasized that the depth of sand-laden wind current and height of stone have much to do with the form of wind-cut surfaces. Bagnold (1942, pp. 36–37) demonstrates that the depth of the sand-laden layer is determined by the height of rebound of grains traveling by saltation. This is controlled by a number of conditions, but it is higher in areas where the ground is mantled with stones than in places where the mantle is solely sand or soil, for the hard stones provide a more resilient rebounding surface. The maximum depth of sand-laden current over sand-covered surfaces is about 18 inches, but over gravel-covered surfaces it may be nearly 4 feet (Bagnold, 1942, pp. 11, 18). A still more important factor is the density distribution of sand grains within the sand-laden current. The mean height of all sand grains traveling by saltation across a sand-covered surface is about 0.9 cm., and most of the grains do not rise above 2 cm. (Bagnold, 1942, p. 63). Above a stone-mantled surface these figures may be more than doubled. One other factor of significance to wind-cutting is the angle of incidence, 10°–16°, between saltating grains and a horizontal surface (Bagnold, 1942, p. 16). These various relations are illustrated in figure 3.

A stone so small that it lies wholly within the dense bottom layer of the sand-laden current, about the lower 4–8 cm. under conditions existing in the Big Horn piedmont, will be cut uniformly, and a facet will be developed. This has been established by Schoewe's experiments (1932, p. 131) and confirmed by Maxson's field observations (1940, p. 735). The distribution of sand grains in this bottom layer may not be entirely uniform, but the greater energy of the grains traveling at higher levels probably compensates for their somewhat fewer number. If the stone rises above the dense bottom layer, its lower part will be cut more vigorously, and a concave face will be formed (fig. 4), at least in the early stages. Sand grains rebounding from the rock surface take longer hops and are more widely dispersed when they hit the rock face at a higher level, and other rebounding grains are deflected to the sides by wind currents flowing laterally around the stone. Thus the number of impacts and therefore the intensity of cutting on the upper part of the stone are reduced, even though some of these grains may have greater striking force because of their longer paths and greater fall. If conditions are stable, the stone can be lowered by continued sandblasting until it finally lies wholly within the zone of maximum and uniform cutting, where the wind-cut face can be reduced to a plane (fig. 4). The face then becomes a facet, and large stones may develop low-angle facets in this way. Lambert (1936, p. 26) states that wind-cut faces are concave when inclined less than 45° with the horizontal and are convex when approaching vertical; but from the above it appears that height of stone is a more important factor, and, as a matter of observation, high-angle faces are not universally convex or low-angle faces universally concave. One should also recognize that the size of the stone is not the important thing but rather the height to which it projects above ground level. A boulder several feet in diameter may be so deeply buried that its upper surface projects only slightly above the ground. Given time, wind-cutting will develop a facet on that boulder.

Steep wind-pitted toes (pl. 3, *E*) probably represent an original steep face at the base of the stone. Cutting on a nearly vertical face is less than on a face inclined 30° from the vertical (Woodworth, 1894, p. 69; Schoewe, 1932, pp. 128–129), presumably because the sand grains rebounding from the vertical face collide with grains moving toward it (Woodworth, 1894, p. 69). The steep toe on a

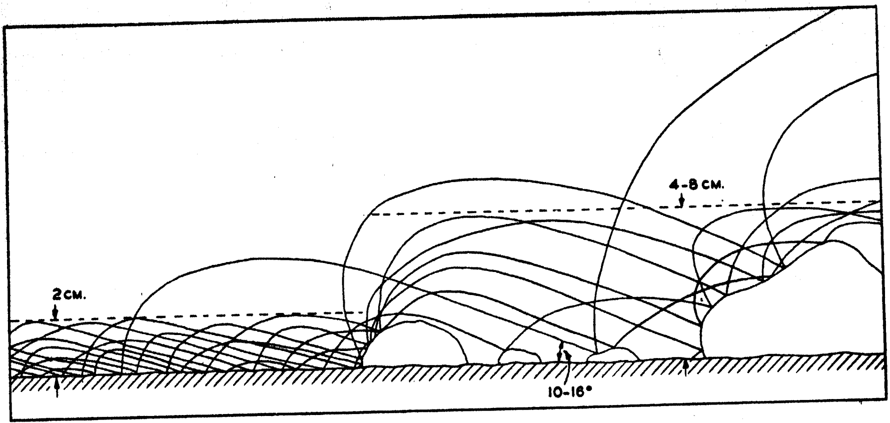


FIG. 3.—Paths of saltating sand grains, showing influence of stones in producing higher rebound. Dashed line marks approximately top of dense bottom layer in the sand-laden wind current. Modified from Bagnold (1942, p. 36).

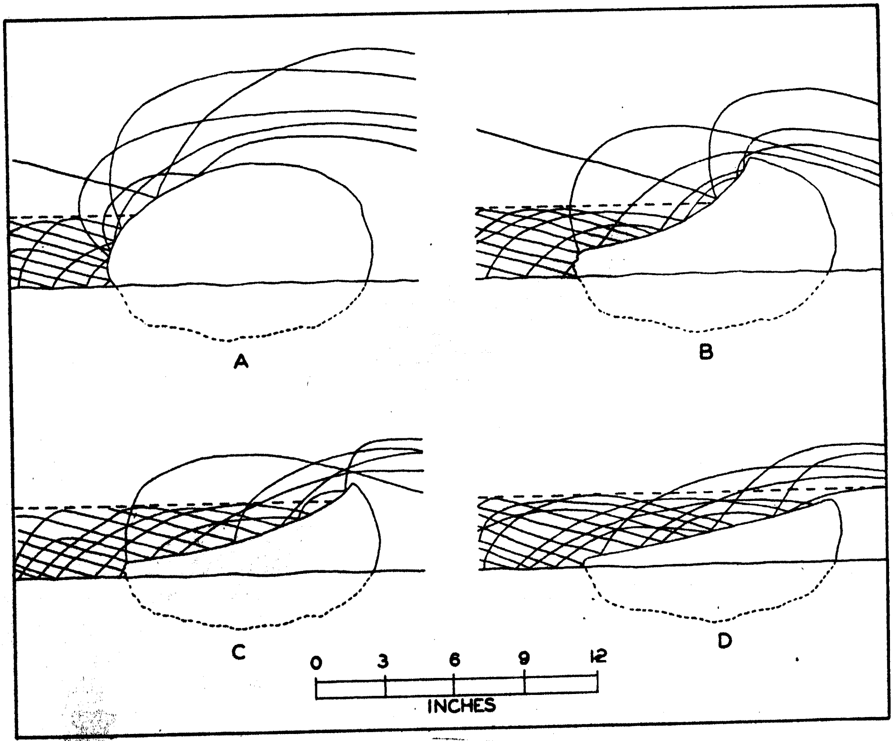


FIG. 4.—Development of concave wind-cut face and eventual reduction to a facet. Sand grains not shown as rebounding from stones are diverted out of the plane of the drawing. Dashed line marks approximate top of dense bottom layer in the sand-laden wind current.

stone would thus not be cut back so quickly as would other more gently inclined surfaces on the same stone.

Wind-cut surfaces on large stones develop a convexity in plan view because of additional cutting by rebounding sand grains deflected to the sides but not clearing the stone on the first hop. If the stone's original surface is curved, this aids lateral diversion. However, the amount of lateral diversion becomes less as the angle of inclination of the surface becomes lower, and very gently inclined faces of considerable width may have almost no convexity in plan view.

Most wind-cut surfaces facing into the wind are inclined 30°–60° from the horizontal. Earlier observations (Bryan, 1931, p. 32; Needham, 1937, p. 33) and experiments (Woodworth, 1894, p. 69; Schoewe, 1932, pp. 128–130) indicate that reduction of faces at angles below 30° proceeds slowly, but cutting does not cease entirely at that angle, as shown by a large number of transverse faces in the Big Horn area with inclinations of 20°–30°. No consistent relation was noted between degree of inclination and lithologic characteristics as recorded by Cloos (1911, p. 59). Stone (1889, p. 420) and Wade (1910, p. 396) state that faces can be planed to horizontal attitude by wind action, but King (1936*a*, p. 206) is skeptical, and Schoewe (1932, p. 129) concludes from his experiments that sand corrasion does not occur on horizontal surfaces.

In this area horizontal wind-cut faces were found on large boulders, on flat platy stones, and on polygonal ventifacts 6–10 inches in diameter. However, it is not possible to assert that any of these faces were reduced to horizontal attitude by wind action; they may have been horizontal joint surfaces or, on the smaller stones, may have been rotated into the horizontal position. Nevertheless, these faces, whatever their origin, have been modified in the horizontal position by wind-cutting; for their flutes and grooves are parallel to the wind. If wind can cut such markings on horizontal faces, there seems no reason why faces cannot be reduced to horizontal attitude by wind action if the stone is stable and the process long continued. Since saltating sand grains strike horizontal surfaces at angles between 10° and 16° (Bagnold, 1942, p. 16), it is easy to see how this cutting could occur. Maxson (1940, pp. 738–739) has observed transverse faces in all positions down to horizontal in Death Valley without being certain that the positions of the faces were due wholly to reduction by wind.

Sharp edges between adjacent faces are considered characteristic of ventifacts (Cailleux, 1942, p. 41), but not all edges need be sharp; for the leading edge of a face is usually blunt, rounded, and pitted, in contrast to the relatively sharp trailing edge, as noted by Heim (1887, p. 384) and Matthes (1934, p. 195).

MULTIPLE FACES

Multiple faces on ventifacts have been attributed to the original shape of the stone and wind-splitting[3] (Heim, 1887, pp. 384–385), to winds from different directions (Stowe, 1873, p. 106; King, 1936*b*, p. 280; Barnes and Parkinson, 1939, p. 670), and to shifting of stones (Woodworth, 1894, pp. 69–70; Bather, 1900, p. 405; Wade, 1910, p. 397; Schoewe, 1932, p. 133; Cailleux, 1942, p. 44; and others). These processes are not mutually exclusive, but in the Big Horn area winds from different directions may be discounted, as evidence is good that the wind blew consistently from one direction. Wind-splitting probably affected some boulders larger than 12–18 inches in diameter and smaller stones of favorable shape and orientation, but it seems clear that wind-splitting cannot explain a large number of faces intersecting in sharp acute edges, cannot account for faces showing great diversity in orientation of grooves and flutes, and probably did not operate effectively on most stones less than 12–18 inches in diameter. Shifting of stones seems the inescapable cause for multiple faces on most Big Horn ventifacts a foot or less in diameter.

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[3] Cloos (1911, p. 54) early took a more moderate view than many Continental Europeans in stating that the principal windward facet on a ventifact will be at right angles to the wind direction.

Some fluted or grooved surfaces show superimposed pitting or a secondary set of flutes or grooves discordant with earlier markings. This would be an expected by-product of stone shifting. If the face had occupied the new position long enough, the older markings would have been destroyed; but in many cases renewed shifting moved the face to a protected position before this happened. A fluted surface showing superimposed pitting must have been rotated so as to face more steeply into the wind. The reverse happened for pitted surfaces with superimposed flutes, and surfaces with two sets of flutes or grooves in different directions have probably experienced oblique tilting or rotation around a vertical axis.

The amount of shifting and the orientation of the axes around which it has occurred can be determined only approximately. Ventifacts with eight or nine narrow facets separated by roughly parallel obtuse edges (pl. 3, *C*) suggest repeated small shifts around a more or less horizontal axis transverse to the wind. Superimposed pits and flutes indicate that this rotation has been downwind in some instances and upwind in others. Pyramidal ventifacts could be produced by rotation around a more or less vertical axis or, if composed of three faces, possibly by rotation around two horizontal axes making a large angle with each other. However, most pyramidal ventifacts in the Big Horn area with the faces limited to the upper half of the stone suggest rotation around a vertical axis.

Shifting of ventifacts has been attributed to undermining by wind scour (Bather, 1900, p. 405; King, 1936*a*, p. 208), overturning by strong winds (Wade, 1910, p. 397; Cailleux, 1942, p. 45), stirring by frost action (Cailleux, 1942, p. 45), and a host of other processes, including rain wash, animals, creep, settling, and solution (Schoewe, 1932, p. 132).

Undermining by wind scour perches stones on tiny pedestals. If sand supply is not too abundant, this scouring occurs on the sides and rear as well as to windward. However, cutting on the windward side is the most intense, and most stones should rotate into the wind about a roughly horizontal transverse axis. Observation of stones on sandy flats in Minnesota and simple experiments with a sand table show this to be the case; but some instances of rotation around horizontal axes parallel or oblique to the wind, because of the shape of the stone or strong lateral undercutting, have been seen. Very strong gusts of wind may cause small perched stones to rotate in a downwind direction (Cailleux, 1942, p. 45); but evidence of such rotation was not observed on the Minnesota sand flats or reproduced experimentally, although markings on Big Horn ventifacts suggest downwind rotation. It is difficult to imagine wind scour causing rotation around a vertical axis, and this difficulty led Barnes and Parkinson (1939, p. 670) to postulate winds from four different directions to explain pyramidal ventifacts from Cambrian sandstone in Texas.

Wind scour may have shifted ventifacts in the Big Horn area, particularly the smaller ones, but wind scouring was by no means universal. The distribution of wind-cut faces on many ventifacts shows clearly that these stones were one-third to two-thirds buried in the ground when cutting occurred. They must have been moved by some other process. Rain wash could shift small stones, but it faces the same difficulty as wind scour with respect to larger, partly buried boulders. The scour and fill attending floodwaters would have incorporated wind-cut stones within the gravel mantle, and that is contrary to field data. In a cold climate freezing and thawing disturb the ground by heaving, solifluction, and related processes, for which Bryan (1946, p. 633) has suggested the inclusive term "congeliturbation." For reasons to be given shortly, it is thought that Big Horn ventifacts were cut during glaciation in the near-by mountains, so congeliturbation and wind-cutting could have been contemporaneous on the Big Horn piedmont. Congeliturbation appears to be the only process capable of shifting stones partly buried in the ground. Furthermore, differential horizontal movements of the mantle rock caused by congeliturbation might have produced the rotation around vertical axes seemingly required by many ventifacts. Denny (1936, p. 336), Bryan (1937, p. 679), and Bryan and Ray (1940, p. 26) describe congeliturbation concurrent with wind-cutting in other areas; and Cailleux (1942, pp. 45, 127) ascribes shifting of many large multifaced ventifacts in Europe to "cryoturbation," an identical process. In the Big Horn region congeliturbation probably outdid wind scour in shifting stones. Whatever the process of shifting, it ceased at about the same time as wind-cutting, for transverse faces on small ventifacts still face northwestward, and the orientation of flutes and grooves shows that many ventifacts have not been moved subsequent to wind-cutting.

CONDITIONS ATTENDING WIND-CUTTING

Ventifacts imply an environment featuring four conditions: (1) adequate but not too abundant supply of a suitable cutting tool, (2) lack of complete vegetative cover, (3) relatively strong winds, and (4) topography permitting free sweep of the wind.

Ventifacts are usually attributed to cutting by wind-driven sand, but Teichert (1939) suggests that wind-blown snow particles at low temperatures may corrade rocks, and Blackwelder (1940, p. 61) summarizes data showing that ice at very low temperatures (−109°.3 F.) may have a hardness approaching 6. However, cutting of Big Horn ventifacts solely by wind-blown ice particles is not favored for the following reasons: (1) Even if temperatures in this area during the glacial ages were consistently −40° to −50° F., ice particles would have had a hardness no greater than 4 (Teichert, 1939, p. 147). (2) Even though ice below −7°.6 F. will not melt by pressure on impact (Barnes, 1928, pp. 86–87), its brittleness at such low temperatures may render it relatively ineffective as a cutting tool. (3) It is yet to be demonstrated that wind-blown ice particles at any temperature can produce ventifacts of the type found here. (4) The ability of ice particles to carve quartz and chert ventifacts at any temperature is highly questionable. (5) Big Horn ventifacts should be more widely distributed if cut by ice particles, for snow was certainly more widely available than sand. (6) Snow is often so abundant, when present, that it buries small stones and protects them from cutting.

For these reasons, sand is the favored cutting tool. It may have come from barren sandy gravel flats along streams, from bare gravel banks on risers between erosion surfaces, or from disintegrating outcrops of Mesozoic and Cenozoic sandstones. The first source appears the most adequate. Some of the sand probably came from the very surface on which the wind-cut stones rested, for wind scour may have been responsible for shifting some stones to allow cutting of multiple faces. Scouring would indicate a limited supply of sand, which is desirable; a too copious sand supply buries the stones before cutting can occur (Schoewe, 1932, p. 131).

The only accumulations of wind-blown sand recognized in this area are small ancient cliff dunes perched along the upper edges of bare north-facing cliffs, from which the detritus composing the dunes has been derived. There are three such occurrences: on top of Moncrief Ridge, on Bald Ridge, and on a high erosion surface 1 mile southeast of the Cross H Ranch near Sand Creek (fig. 1). The dunes on Moncrief Ridge contain remnants of four soil profiles, suggesting four periods of deposition separated by episodes of weathering and vegetative growth. This scarcity of sand accumulations is not a serious problem; for other regions of active wind-cutting are devoid of accumulated sand and the total amount required for cutting may not be great. In the Big Horn area the wind blew transversely across the piedmont slope, so that sand picked up from the barren flats along one stream could have been carried by the wind and dumped into the next stream, provided that the path was clear of topographic or vegetative obstructions. At times barren flats probably extended from one stream to the next, as sketched (fig. 5), so wind-blown sand could have been disposed of in this manner. It seems likely that some accumulation of sand did occur, but such deposits, if not removed by subsequent erosion, have been so modified by weathering, wash, and mass movement as to have escaped recognition.

Vegetation is a powerful factor in restricting wind-cutting; even the present sparse cover of grass and brush is enough to prevent sandblasting. However, complete absence of vegetation is not necessary, for wind-cutting occurs in some desert areas supporting a sparse growth of widely scattered bushes. The problem is to find some means of producing and maintaining broad areas almost entirely barren of vegation.

Vegetative cover may be reduced or destroyed by climatic variation or geological change; but climatic variation alone, without accompanying geological change, probably would not produce optimum conditions for sandblasting. A change to a cold climate would actually enhance the vegetative cover by decreasing evaporation (Ray, 1940, p. 1912) and, if severe enough, by producing a frozen subsoil which holds moisture near the surface. A severe warm-dry climate could reduce vegetation to the extent that sandblasting might occur; but such a climate should reduce vegetation all over the piedmont, thus permitting synchronous cutting on high and low erosion surfaces. This is contrary to field fact, for ventifacts on high surfaces are old and those on low surfaces are young.

Geological changes most likely to have destroyed vegetation in the Big Horn piedmont are dissection and deposition. In areas of shallow ground water, dissection can alter vegetation by lowering the ground-water table (Cooper, 1935, pp. 88, 95, 102, 108), but the efficacy of this process in a region of relatively deep ground water, such as the Big Horn piedmont, is open to question.

Deposition appears much more promising, especially the somewhat catastrophic widespread scour and fill produced by floods. Broad river flats or piedmont surfaces mantled by recently deposited sand and gravel are ideal for sandblasting on three counts: (1) they are bare of vegetation, (2) their flatness permits free sweep of the wind, and (3) they provide both a source of sand to do the cutting and stones to be cut. Encroachment of new vegetation on such flats would probably be delayed long enough by vigorous sandblasting to permit cutting of ventifacts.

Deposition on all the erosion surfaces of the Big Horn piedmont is attested by their thin veneer of stream gravel, but this deposition was primarily of the cut-and-fill variety, much like that along the present streams. Unless climatic conditions and stream behavior were much different at earlier periods, such cut-and-fill produced only narrow strips of raw debris, closely confined on both sides by vegetation. The winds of this area, transverse to such narrow strips, could not have obtained sufficient fetch to cut ventifacts. Broader areas must have been inundated by a more catastrophic process, probably floods, which spread over erosion surfaces lying close to stream

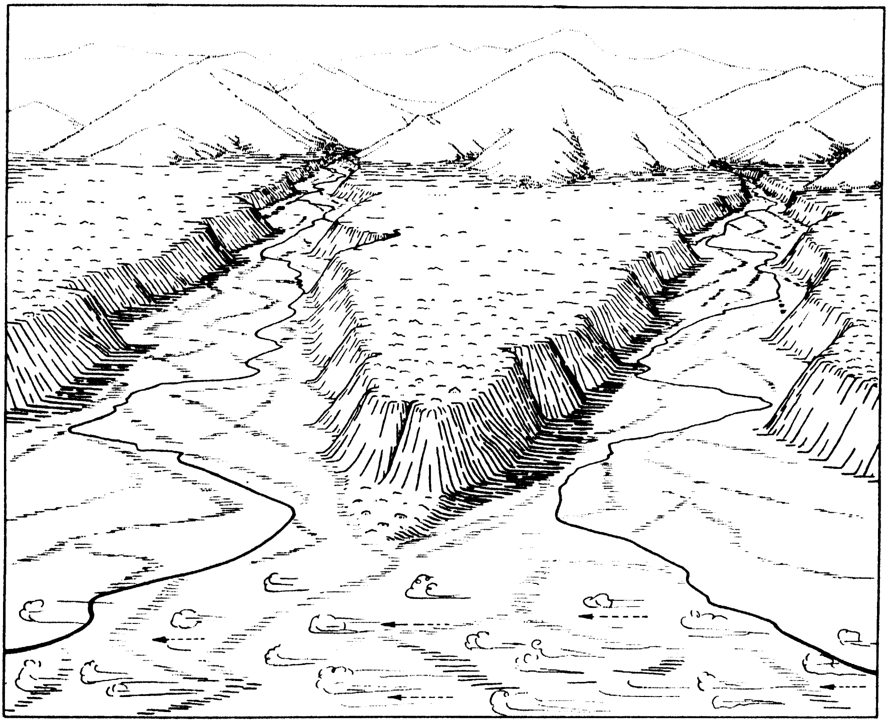

FIG. 5.—Topographic situation in Big Horn piedmont during wind-cutting. Old ventifacts on high vegetation-covered erosion surfaces protected from sand-blasting under way in foreground. Arrows indicate wind direction.

grade. Those parts of the piedmont close to the mountains across which the streams flowed in incised courses would not have been affected, but, farther out, wide areas probably extending from one stream to the next could have been inundated (fig. 5). At earlier stages, when the higher surfaces were at stream grade, they, too, could have been inundated.

Floods on the Big Horn piedmont could have been produced by either warm-dry or cold climatic conditions. Floods resulting from a cold climate would consist of meltwater from the glaciers in the Big Horn Mountains, while those produced by a warm-dry climate would be of the cloudburst variety characteristic of semiarid and arid regions. In either case the total amount of detritus carried out from the mountains need not have been exceptional. The floods may simply have re-worked debris that already lay on the piedmont surface. This would have been the case particularly for glacial floods, as most of the outwash debris from the mountain glaciers is trapped in shallow valleys on the subsummit upland (fig. 6) and no valley trains extend to the eastern base of the range (Darton, 1906, pl. 26). The lack of thick gravels on the piedmont erosion surfaces need not rule out the floodwater hypothesis.

The major problem is to choose between a warm-dry and a cold climate. The latter is favored, for, as will be shown shortly, wind-cutting took place at several different times and the climatic conditions favorable to sandblasting must have occurred repeatedly. The various Pleistocene glacial stages and Wisconsin stadials provide ample cold periods, but warm-dry periods are less abundant. The post-Wisconsin optimum, 8,000–4,000 years ago, was a period warmer and drier than the present and perhaps favorable to wind-cutting; but one must go back to pre-Wisconsin interglacial stages to find other suitable warm-dry periods. The various interstadials of the Wisconsin were probably cooler and possibly more humid than the present. Thus all but the youngest Big Horn ventifacts would have to be pre-Wisconsin if formed during warm-dry periods, and this seems highly unlikely. On the other hand, not only are cold periods numerous, but they are known to have been accompanied by strong winds, and frozen ground is particularly favorable for movement of sand. In so far as

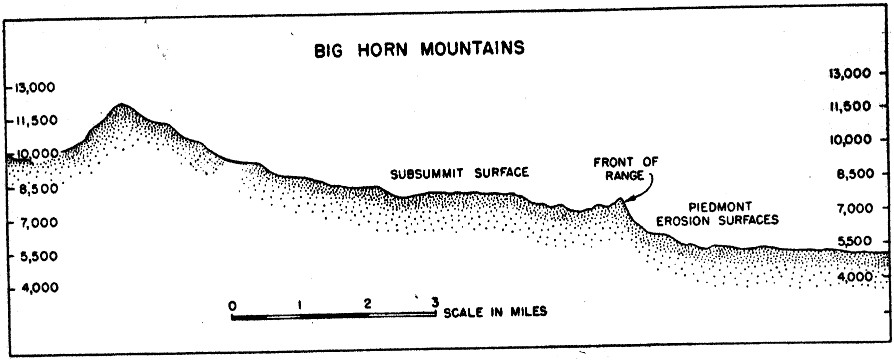

FIG. 6.—Topographic profile of east half of Big Horn Mountains

congeliturbation (Bryan, 1946, p. 633) may have been responsible for the turning of ventifacts with multiple faces, cold conditions are further indicated.

For these reasons it is maintained that the Big Horn ventifacts were formed almost wholly on sand and gravel flats inundated by floods of meltwater from near-by mountain glaciers and exposed to the strong winds of a periglacial environment. Antevs (1928), Hobbs (1942; 1943, pp. 369–378), and Belknap (1941, p. 235) have emphasized the aspects of such an environment which are highly favorable to wind-cutting. At each period large areas of the Big Horn piedmont were not inundated and were not suitable for wind-cutting, and this accounts for the patchy distribution of ventifacts, as well as the different ages of ventifacts on erosion surfaces of different levels.

AGE OF VENTIFACTS

The antiquity of these ventifacts is demonstrated by the following facts: (1) they are located on brush- and grass-covered flats (pl. 2, *A*), across which no sand could be blown at present with sufficient force to cut the stones; (2) wind-cut faces are partly, and in some cases wholly, buried in a deposit of fine sheet wash; (3) exposed faces are commonly encrusted with lichen; and (4) many ventifacts on the higher erosion surfaces are extensively weathered (pl. 1, *B*).

Ventifacts have been found on erosion surfaces of at least four distinct levels. For example, north of Clear Creek just west of Buffalo (fig. 1), ventifacts lie on surfaces 25, 115, 265, and 285 feet above present stream grade. South of French Creek they occupy surfaces at four different levels 25–325 feet above stream grade. This relation in itself suggests that not all the ventifacts are of the same age, a conclusion amply borne out by differences in degree of weathering.

The most-abundant and best-preserved ventifacts are on the lowest surfaces, Alden's (1932, pl. 1, east half) No. 3-bench group, where rocks of nearly all types except limestone and incoherent sandstone exhibit evidence of wind-cutting. Here easily weathered rocks, such as coarse granite, diabase, and metamorphics rich in ferromagnesian minerals, have well-preserved wind-cut faces. On the highest erosion surfaces, the No. 2-bench group, ventifacts are sparse and badly weathered, and wind-cut surfaces are preserved only in vestiges (pl. 1, *B*) on the most resistant rocks, usually those of dense texture and siliceous composition. Transitions between the two extremes of abundant fresh ventifacts on a wide variety of rocks and sparse deteriorated ventifacts on predominantly siliceous rocks are found on erosion surfaces at intermediate levels.

Four separate periods of wind-cutting cannot be established solely from the degree of preservation of ventifacts, although it is frequently possible to say, on the basis of weathering, that the ventifacts at one place are younger or older or about the same age as those of another locality. The fact that ventifacts showing differing degrees of weathering rest on erosion surfaces of four different levels is held to be ample proof that at least four separate periods of wind-cutting occurred. The remnants of four soil profiles in deposits of wind-blown sand on Moncrief Ridge may also be significant in this connection.

If the four separate periods are accepted, then it must be assumed that wind-cutting on an erosion surface occurred before it was dissected. That this need not be universally true is suggested by Wentworth and Dickey (1935, p. 101) and demonstrated on Moncrief Ridge, a high piedmont ridge between North Piney and Little Goose creeks (fig. 1) at an altitude of 6,500 feet. Along the north edge of Moncrief Ridge are small cliff dunes, composed of disintegrated debris blown up from a bare cliff of coarse granitic gravel which composes the north face of the ridge. Between the dunes and the edge of the cliff is a number of relatively fresh ventifacts. Moncrief Ridge is much higher and older than any of the erosion surfaces upon which extensively weathered and presumably older ventifacts have been found; thus ventifacts can be cut along the upwind edge of an erosion surface after it has been dissected. However, such action is limited to the edge of the surface above a bare cliff, and ventifacts lying more than 100 or 200 feet back from the edge must have been cut before the surface was extensively dissected. Vegetation is a major factor in this connection; for an erosion surface will be bare of vegetation only during and for a short time after its gravel cover has been spread or reworked by floods. By the time that dissection occurs, enough vegetation should have developed to prevent much movement of sand. Wind-cutting on dissected surfaces is further handicapped by topographic isolation from river floodplains, the major source of sand (fig. 5). Therefore, the age of the ventifacts, with few exceptions, would appear to be the same as the age of the erosion surface upon which they lie.

The erosion surfaces of the Big Horn piedmont can hardly be older than Pleistocene (Alden, 1932, pp. 44, 59), and some relation with the Pleistocene glacial succession would be expected. Except for the lowest terrace, these surfaces are cut on rock; they are not gravel fills, and one cannot assert that each surface corresponds to a phase of glacial outwash from the near-by mountains, as is partly the case farther south (Bryan and Ray, 1940, p. 15). Furthermore, the local topographic situation precludes any attempt to trace the levels of ventifact-bearing surfaces back to the glaciated areas of the Big Horn Mountains. The backbone of this range is a series of high peaks flanked by a broad subsummit surface of gentle relief at 7,500–9,000 feet (fig. 6). This subsummit surface is separated from the piedmont by the abrupt east face of the range, composed of sharply upturned Paleozoic beds. Glaciers from the high peaks moved onto the subsummit surface, but nowhere did they reach the east base of the range.[4] Streams from the glaciated areas pass out of the range in deep narrow rock gorges through which successive outwash trains cannot be traced. Thus it is impossible to relate the piedmont erosion surfaces directly to the two stages of Pleistocene glaciation recognized in the mountains (Salisbury and Blackwelder, 1903, pp. 216–220).

On the basis of data from other areas, Alden (1932, p. 59) states that the No. 3 bench was completed as early as the oncoming of the Wisconsin, which would mean the Tazewell or some later substage of the Wisconsin as the sequence is now interpreted. He (Alden, 1932, p. vii) correlates the No. 2 bench with the Circle terraces, which were presumably at stream level during the Bull Lake (Iowan) substage of the Wisconsin (Blackwelder, 1915, p. 316; Fryxell, 1930, pp. 39–42; Horberg, 1938, p. 75). Horberg (1940, p. 292), however, states that the No. 3 terrace (bench) along Yellowstone Valley was completed prior to the earliest Wisconsin (Iowan) glaciation. If correct, this would make the No. 2 bench pre-Iowan, and Alden (1932, p. 32) has found evidence farther north indicating that it may be as old as Kansan or Nebraskan. Antevs (1945, pp. 24–25) suggests that Bull Lake glaciers may have been pre-Wisconsin, which, coupled with Horberg's interpretation, would make both the No. 2 and the No. 3 benches pre-Wisconsin. This seems unlikely; but, in view of the above inconsistencies, the only conclusion possible is that the erosion surfaces east of the Big Horns are Pleistocene, the lower ones probably Wisconsin, and the higher ones possibly pre-Wisconsin.

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[4] Alden (1932, p. 43) has suggested that great piedmont gravel ridges, such as Moncrief, Bald, and North (fig. 1), along the eastern base of the range are of Pleistocene glacial origin. This view is not accepted here, and the origin of these gravel ridges is the subject of a separate paper (Sharp, 1948).

For reasons already given, optimum conditions for wind-cutting in this area are thought to have occurred during times of glaciation in the Big Horn Mountains. However, wind-cutting occurred only upon an erosion surface essentially at stream grade. At times, streams of glacial meltwater from the mountains probably flowed across the piedmont in incised channels, so that broad, barren flats were not developed and no ventifacts were formed. Thus each phase of glaciation in the mountains need not have given rise to a period of wind-cutting on the piedmont; ventifacts were formed only when a piedmont erosion surface lay close to stream grade during a phase of mountain glaciation.

Certainly, the recognition of four periods of wind-cutting should not be interpreted as indicating correlation with the four major stages of Pleistocene glaciation. It seems likely that more than one of the periods of wind-cutting occurred within the Wisconsin; for at least two substages of Wisconsin glaciation are widely recognized in western mountains (Blackwelder, 1931, p. 918; Antevs 1945, Table II, p. 24), and Bryan and Ray (1940, pp. 27–35) have distinguished five substages of the Wisconsin farther south in the Rockies. The oldest ventifacts on the highest surfaces may well be pre-Wisconsin, but it is not possible to assign them to any specific pre-Wisconsin stage. Powers (1936, pp. 216–219) recognizes pre-Wisconsin wind-cutting on terrace boulders in the southern Rockies.

PLEISTOCENE WIND DIRECTION

In determining ancient wind directions from ventifacts (Powers, 1936, pp. 215, 218; Denny, 1941, p. 257; Dow, 1940, p. 476; Cailleux, 1942, p. 73; and others), the major problem is to show that the stones have not been moved since the wind-cutting. In the Big Horn region large boulders with wind-cut faces on only one side have clearly not been shifted, since the faces on a number of such boulders at one locality all face the same direction (pl. 2, *A*). Usually, it can be shown by reference to these large stones that many of the smaller ventifacts have not been moved since the last episode of wind-cutting, and they can also be used for measuring wind direction.

Measurements of wind direction are most reliable when made with reference to the orientation both of a face and of the grooves, flutes, or pits on its surface. Pitted surfaces faced more or less directly into the wind; grooved or fluted faces may have been transverse or oblique to the wind. If these markings are parallel to the dip line of the surface, it faced directly into the wind. If they are at an angle to the dip line, the face was oblique to the wind and should not be used. The directions of flutes and grooves across oblique faces are also unreliable, as orientation of the faces controlled to some degree the direction of air movement. However, flutes or grooves on horizontal or gently inclined surfaces can be used with reasonable reliability. In this work, eight to twenty readings were made on separate stones at each locality and were averaged. Consistent results were obtained by following this practice and by observing the relations between faces and surface markings outlined above.

TABLE 1

TRUE WIND VELOCITIES AT SHERIDAN, WYOMING, IN MILES PER HOUR (WYLAND, 1945, p. 3)

	Average Hourly Velocity	Prevailing Direction	Highest Velocity*	Direction of Highest Velocity
Length of record in years	33	33	36	36
January	5.1	NW.	57	NW.
February	5.2	NW.	45	NW.
March	6.4	NW.	52	NW.
April	7.3	NW.	47	NW.
May	7.2	NW.	47	NW.
June	5.7	NW.	42	NW.
July	4.9	NW.	49	SW.
August	4.7	NW.	58	NW.
September	4.8	NW.	59	NW.
October	5.3	NW.	49	NW.
November	4.7	NW.	47	NW.
December	4.9	NW.	47	NW.
Average for year	5.5	NW.	53	NW.

* Maintained for a period of 5 minutes or more. Usually velocities one-half to two-thirds these values were maintained for several hours before and after the peak maximum of the storm and from the same general direction.

The map (fig. 1) shows wind directions thus determined at twenty-nine localities. Reliable measurements could not be made at all ventifact localities because of the scarcity of wind-cut stones, high degree of weathering, or lack of large, stable wind-cut boulders.

It is clear that at all periods of wind-cutting the prevailing winds were from the northwest, ranging from N. 10° W. to N. 60° W., with a concentration in a N. 20° to 30° W. direction and an average for all readings of N. 29° W. Ventifacts along the base of the Colorado Front Range also indicate a northwest wind (Stone, 1889, p. 421).

Winds in the Big Horn region had a slightly stronger northern component away from the mountain front, although this tendency was occasionally outweighed by local topographic conditions causing deflection of ground winds. Close to the mountains the winds were roughly parallel with the mountain front, and in the northern part, where the range swings more to the west, winds show a similar change (fig. 1).

Averaged readings from ventifacts on erosion surfaces at different levels are within 2° or 3° and indicate no significant difference in wind direction during the different periods of wind-cutting. No evidence was found of strong winds sweeping down the canyons from the mountain glaciers, as reported by Blackwelder (1929*b*, pp. 256–257; 1931, p. 878) in the Sierra Nevada, by Bryan and Ray (1940, pp. 30, 34) and Ray (1940, pp. 1857–1859) in the southern Rockies, and by Odell (1925, p. 306) in the Himalayas. Presumably, the topographic conditions already outlined precluded such winds.

Comparison with modern wind records from the Sheridan area (table 1) indicates that Pleistocene and modern wind directions were essentially the same and that orographic control was dominant then as now. Local mountain glaciers seem to have had no effect, nor can it be shown that the continental ice sheet, 210–250 miles north, exerted any perceptible influence.

CONCLUSIONS

1. True wind-cut facets are formed only when that part of the stone projecting above the level of the ground lies

wholly within the dense bottom layer of the sand-laden wind current, about the lower 4–8 cm. under conditions ruling in the Big Horn piedmont. Stones projecting to greater height tend to develop various types of curved faces, which may be reduced to facets when the height of the stone has been sufficiently lowered.

2. Wind-cut faces making angles greater than 55° with the wind are pitted. Faces making angles less than 55° with the wind are fluted or grooved.

3. Flutes on high-angle faces are short and deep; on low-angle faces they are long and shallow. Grooves are more abundant on low-angle faces, but the two forms are not mutually exclusive.

4. Pitting is closely controlled by mineral hardness and rock structure. Flutes and grooves are independent of such features unless structure happens to parallel wind direction.

5. Wind-cut faces rotated into different positions may have flutes, grooves, or pits superimposed on earlier markings of different nature or of different orientation.

6. Cutting of Big Horn ventifacts is attributed to wind-blown sand and not to wind-blown ice particles of low temperature and exceptional hardness.

7. Cutting has occurred on horizontal surfaces, but it cannot be shown that faces are reduced to horizontal attitude by wind action, although there is no apparent reason why they should not be.

8. Multiple faces are attributed to stone-shifting caused partly by wind scour and more generally by congeliturbation, which appears to be the only process likely to produce the rotation around a vertical axis that seems to be required in many instances.

9. Four separate periods of wind-cutting are indicated by ventifacts showing different degrees of weathering and resting on erosion surfaces at four separate levels, 25–325 feet above stream grade.

10. Conditions for sandblasting are shown to have been at an optimum during episodes of glaciation in the near-by mountains when floods of meltwater inundated parts of the Big Horn piedmont and strong winds prevailed.

11. Ventifacts on the lower surfaces are probably Wisconsin, and those on higher surfaces are possibly pre-Wisconsin, but no assignment to specific stages or substages of the Pleistocene glacial succession is possible.

12. Pleistocene wind directions can be measured by references to wind-cut faces, flutes, grooves, and pits on large boulders. The average of wind directions so determined in twenty-nine separate localities is N. 29° W., with extremes ranging from N. 10° to 60° W. and no discernible difference between the four separate periods of cutting.

13. Modern winds are also consistently from the northwest, showing that local orographic conditions have controlled Pleistocene winds as well as those of the present. Glaciers in the mountains or the continental ice sheet 210–250 miles north seemingly had little influence on wind directions.

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PERIGLACIAL FEATURES IN THE DRIFTLESS AREA OF SOUTHERN WISCONSIN[1]

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ABSTRACT

At several places in the Driftless Area there occur anomalous surface features not readily explicable on the basis of processes now active. In the Baraboo area, blocks of quartzite from the pre-Cambrian bedrock have accumulated in block streams, block fields, rubble zones, and talus, on slopes ranging from 5° to 36°. Locally, small valleys have been choked by the rubble. In the Blue Mound area, block streams, block fields, and isolated rock masses, derived from the Niagara chert which caps the "mound," are found at distances of up to nearly 1 mile from the parent ledges and on slopes down to slightly less than 4°. At both places the block accumulations are either surrounded or partially overgrown by vegetation and show the effects of long-continued weathering. Nowhere is there any indication of appreciable movement at the present time. The features in question are "fossil" forms, now in a state of stagnation and decay. They are best explained as products of periglacial frost weathering, solifluction, and frost heave during Wisconsin time.

INTRODUCTION

In 1897, T. C. Chamberlin, commenting on certain anomalous features in the Driftless Area, wrote: "It must be remembered that as a result of the excessive superficial thawing and freezing incident to glacial-border conditions, the facilities for landslides, bodily creeps, and similar modes of movement reached an extraordinary degree of development." At that time, however, there was comparatively little basis for particularizing further on the effects of the processes referred to. Since that time extended studies in arctic and subarctic regions have added greatly to our knowledge of frost phenomena, and the concept of a frost climate in former periglacial or near-glacial zones has taken definite form. In the nonglaciated areas of northern Germany and surrounding countries there has been general recognition of surficial deposits and "fossil" land forms unrelated to present-day processes and dating back to the ice age. In North America it was not until Bryan's introduction of European concepts, in 1928, that any attention was given to similar phenomena. Since then, periglacial features have been studied at scattered localities by a comparatively small group of workers.[2]

The present study was undertaken primarily to add to the limited knowledge of periglacial phenomena in the United States. The Driftless Area was selected as particularly promising for such a study because of its proximity to glacial lobes both to the east and to the west and because its varied topography and geology seemed likely to have provided opportunity for the formation and preservation of periglacial features in at least some places.

Field work was carried on during the summers of 1940 and 1941, and the time totaled between 3 and 4 weeks. It was mainly of reconnaissance character, being directed, first, to the search for significant localities and, second, to the more detailed examination of such localities as seemed most promising. The localities studied are shown on the map, figure 1.

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[1] Manuscript received December 23, 1948.

[2] A more comprehensive review of the study of periglacial phenomena is being published elsewhere as one section on a report by the Committee on Interrelations of Pleistocene Research, of the National Research Council.

PREVIOUS STUDIES IN THE AREA

Previous studies have been concerned primarily with general geology and physiography. In 1886 the broader features of the area were outlined by Chamberlin and Salisbury. In 1918 the glacial geology of adjoining areas was described by Alden. Regional physiography was summarized by Martin in 1932. In 1941 an excellent physiographic diagram of the area described in this paper was published by Trewartha and Smith. Other studies of a more local nature are referred to in later sections of this paper.

The occurrence of seemingly anomalous deposits in the area was noted by Sardeson and by Squier, in 1897, and was attributed to local glaciation. The localities described by Sardeson are discussed in this paper, but those mentioned by Squier could not be located from his rather indefinite descriptions, within the time available.

THE BARABOO AREA

GENERAL RELATIONS

The location of the Baraboo area (area *A*)[3] is shown in figure 1, and the generalized topography is shown in figure 2. The salient topographic features are the North and South Baraboo ranges (Martin, 1932, pp. 55–57), which represent the two flanks of an asymmetric syncline in pre-Cambrian rocks. The South Range is a broad, flattish upland, strongly dissected along the margins and transected by the gap in which Devil's Lake (locality *A-4*) is located. The North Range is narrower and discontinuous, being cut by several stream gaps. Both ranges are held up by the massive Baraboo quartzite; dips, however, are much steeper in the North

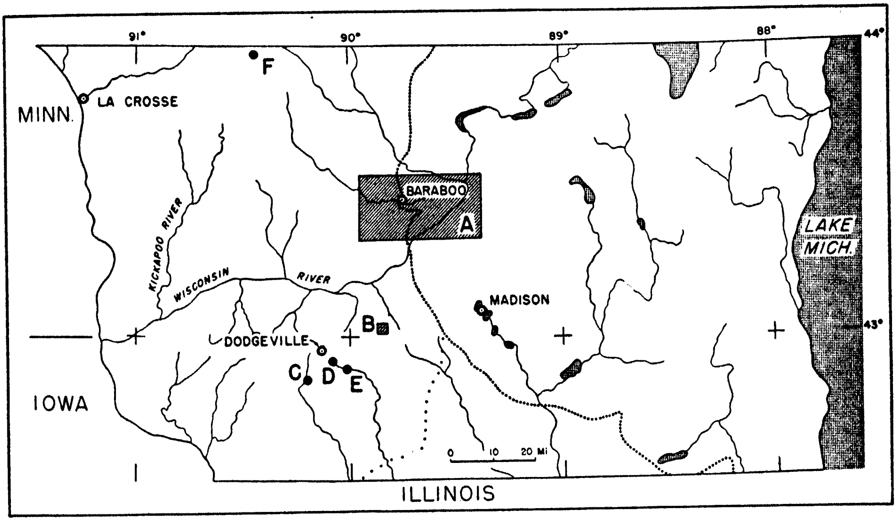

FIG. 1.—Sketch map of Driftless Area, showing location of places described in the text. Each locality is designated by letter. Detailed maps of localities *A* and *B* are found in figs. 2 and 3, respectively. The heavy dotted line marks the boundary of Wisconsin drift, and the light dotted line the boundary of the Illinoian drift.

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[3] In this paper the areas and localities are referred to by letter, as indicated in fig. 1; individual localities within area *A* are designated by numerals following the letter and are numbered as shown in fig. 2.

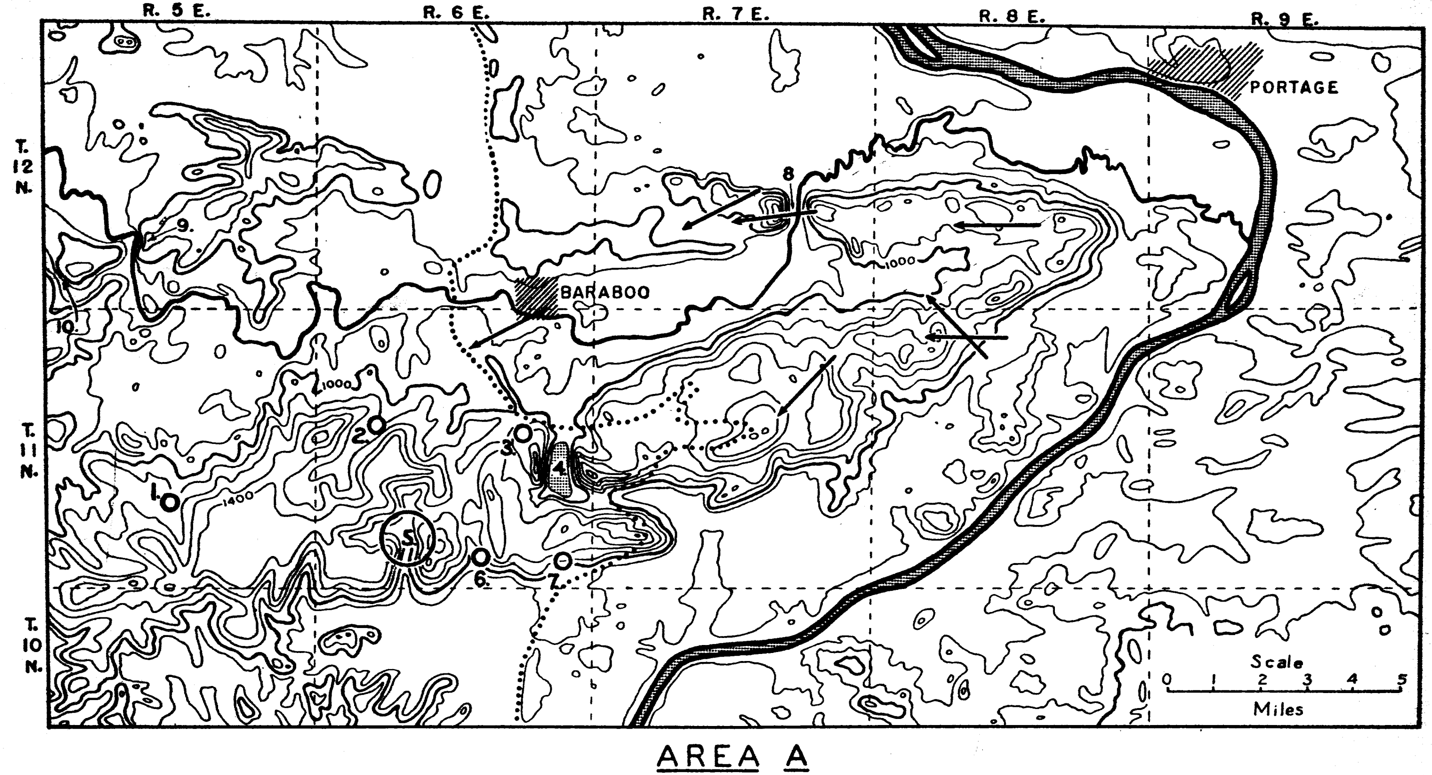

FIG. 2.—Generalized topographic map of the Baraboo area, compiled from the Baraboo, Denzer, Portage, and the Dells topographic sheets. The glacial border is shown by a dotted line, and glacial striations by arrows (both according to Alden, 1918). Localities described in the text are designated by numerals; all except *8* are sites for periglacial features. Contour interval is 100 feet.

Range than in the South Range (Weidman, 1904; Leith, 1935). The physiographic history of the area is complex (Thwaites, 1935), with glaciation during the Cary substage of Wisconsin time as the last major event; the ice advanced from the east (Alden, 1918).

Within the Baraboo area the features attributed to periglacial processes are of three main types: (1) stabilized talus; (2) block concentrations and block-strewn slopes; and (3) choked valleys and block cascades. The distribution and interpretation of each of these is set forth below.

TALUS DEPOSITS

The principal occurrence of talus is in the vicinity of Devil's Lake (locality *A-4*). According to Salisbury and Atwood (1900, p. 67), talus was present originally also in the gorge north of Ableman (locality *A-9*), but, since the time of their writing, it has been largely removed by quarrying operations. Block accumulations transitional between talus and block fields are found along the gorge of Narrows Creek, a few miles west of Ableman (locality *A-10*). Of these localities, only the one at Devil's Lake warrants detailed consideration.

Geologic setting of Devil's Lake.—Devil's Lake occupies the north-south portion of a deep trough cut through the South Range.[4] The bordering bluffs are steep and rocky, with a maximum relief of about 500 feet. The maximum depth of the lake is 43 feet (Juday, 1914, p. 33). Between the lake bottom and the bedrock floor of the trough are more than 200 feet of glacial drift. From the south end of the lake the topographic sag swings eastward and descends to the adjacent lowland. As it crosses the glacial border, the slopes lose their rugged appearance.

The trough in which Devil's Lake lies represents a pre-Wisconsin gorge cut through the South Range by the ancestral Wisconsin River. When the advancing ice reached its maximum extension, this gorge was blocked at both ends, forming a lake with water level some 230 feet higher than the present lake level (Trowbridge, 1917). When the ice retreated, the lake was lowered through successive stages to its present level. The Wisconsin River established a new course, and its former gorge, dammed at both ends by deltaic moraines, was left as a lake basin.

Description of the talus.—Talus occurs all along the unglaciated portion of the Devil's Lake trough. On the bluffs bordering the lake itself, the appearance is essentially the same on both sides. The talus is virtually continuous laterally, except where interrupted by dipping ledges of quartzite. It is partly covered, however, by irregular, discontinuous stands of trees, which occur from top to bottom. Where the topographic trough swings east from the south end of the lake, the appearance of the two slopes changes. On the south-facing slope the talus attains maximum height and continuity of exposure (pl. 1, *A*).[5] On the north-facing slope, however, all except the uppermost part of the talus is covered by forest; on this side the slope is somewhat less steep also.

The talus is composed of heterogeneous, angular, irregular blocks, more or less firmly wedged together. The blocks range up to more than 6 feet in length.

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[4] A detailed topographic map of the locality is found in the works of Martin (1932, p. 122) and of Thwaites (1935, fig. 244). The area is shown also on aerial photos WR-13-1071 to 1073, inclusive (dated 1937), of the U.S. Department of Agriculture, P.M.A.

[5] For other views of the talus see Salisbury and Atwood (1900, pls. 6, 22, and 24), and Martin (1932, pl. 10).

There is no marked vertical zoning of the larger blocks. Except for occasional erratic boulders believed to have been ice-rafted (Salisbury and Atwood, 1900, p. 133), the rock is all quartzite. A preponderance of the blocks are mottled by lichens and discolored by weathering. Surmounting the talus at many places are near-vertical rock bluffs up to several tens of feet in height, with a rough and ragged appearance. The maximum height of the talus is roughly 300 feet. The maximum slope is slightly above 36°.

The basal portions of the talus show various modifications. Along the east side of Devil's Lake, oversteepening resulted from the excavation of the railroad cut. Along the west side of the lake, the slope of the talus is somewhat flattened near the water's edge. Near the southeastern end of the lake, where the talus comes in contact with water-laid drift, the contact is marked locally by discontinuous, elongate depressions up to several feet in depth. These seem best explained as due to settling of the sediment into interstitial openings in the buried talus (Thwaites, 1935, p. 404).

The hydrographic map of Devil's Lake (Juday, 1914, map 8) suggests that the talus extends to 30 or more feet below water level. Additional information on the subsurface extent of the talus is provided by a well at the shelter house near the south end of the lake.[6] The well is located less than 350 feet from the foot of

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[6] F. T. Thwaites, in letter dated December 8, 1948.

PLATE 1

A, Talus just beyond Wisconsin drift border, near southeast side of Devil's Lake (locality *A-4*), looking northeast.

B, South slope of Baraboo Narrows (locality *A-8*), within the glaciated area, showing virtual absence of talus.

C, Scattered blocks of massive chert on east slope of west Blue Mound (locality *B*), looking southwest; the slope is between 3° and 4°.

PLATE 2

Chert blocks on slopes of west Blue Mound (locality *B*).

A, Close-up view of one of the larger chert blocks on the eastern slope, showing disintegration and breakdown in place. The stick gives scale, being slightly more than 5 feet long; the block itself is approximately 20 feet long.

B, Block stream on north slope.

C, Perched block in same locality as *B* above, indicating long-continued stability of the slope. The stick gives scale, being slightly more than 5 feet long.

PLATE 3

Sloping block field, 1.3 miles south of Devils Lake (locality *A-7*).

A, Surface of the block field, showing jumbled appearance and lichen-covered condition of the blocks. The rock is quartzite.

B, Excavation at the base of the block field, showing earth-free character to a depth of 10 feet.

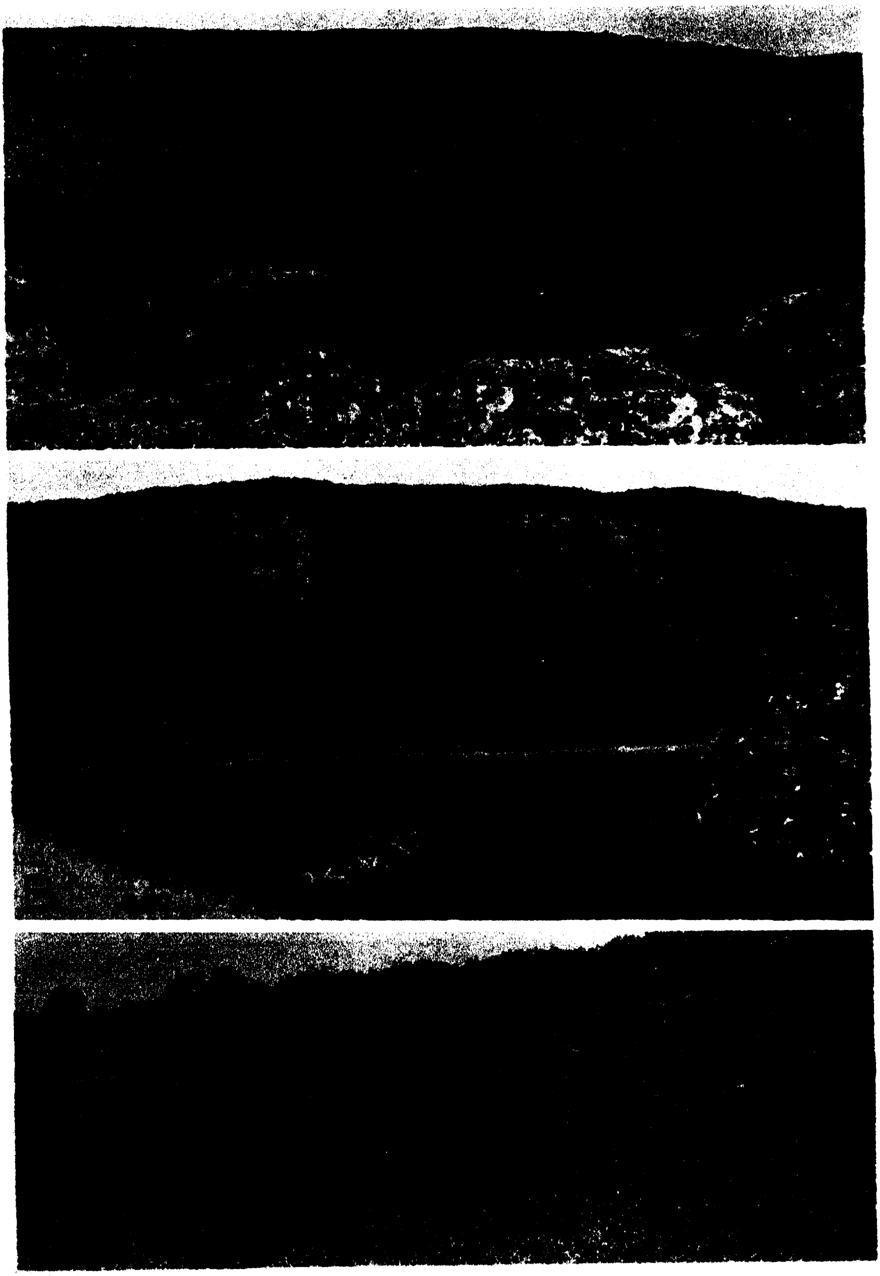
C, Excavation in vegetation-covered portion of the blocky slope, adjoining *B* above, showing earthy matrix between blocks.

PLATE 4

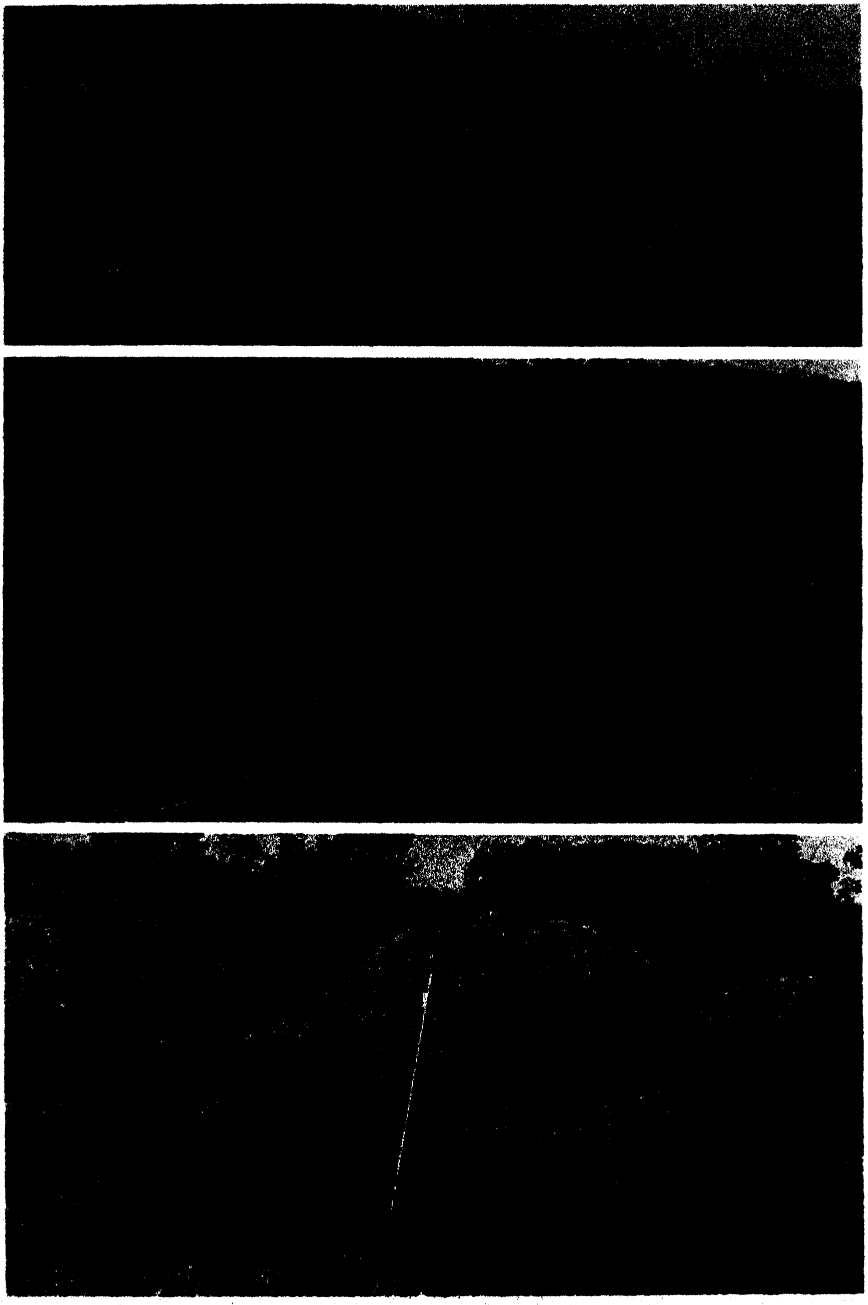
A, Band of blocks marking position of hidden subsurface streamlet along valley flat of Pine Creek, at locality *A-2*. The valley has been choked by detritus, and normal surface drainage has not yet been restored.

B, Transverse block bench across valley of Pine Creek (locality *A-2*), downvalley from *A* above.

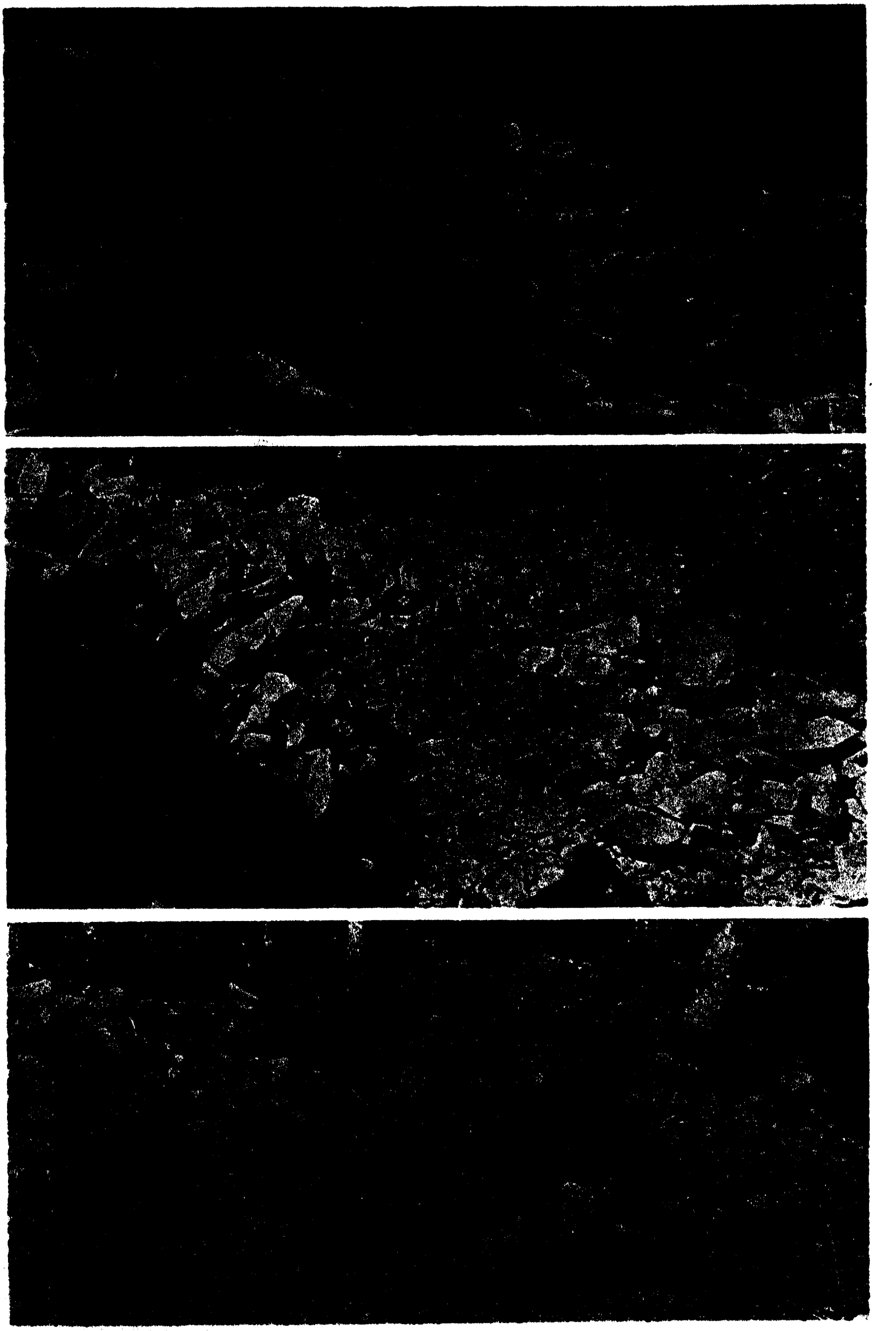
C, Block stream in Baxter Hollow (locality *A-5*); length is about 400 feet, maximum width about 75 feet, and slope about 5°.



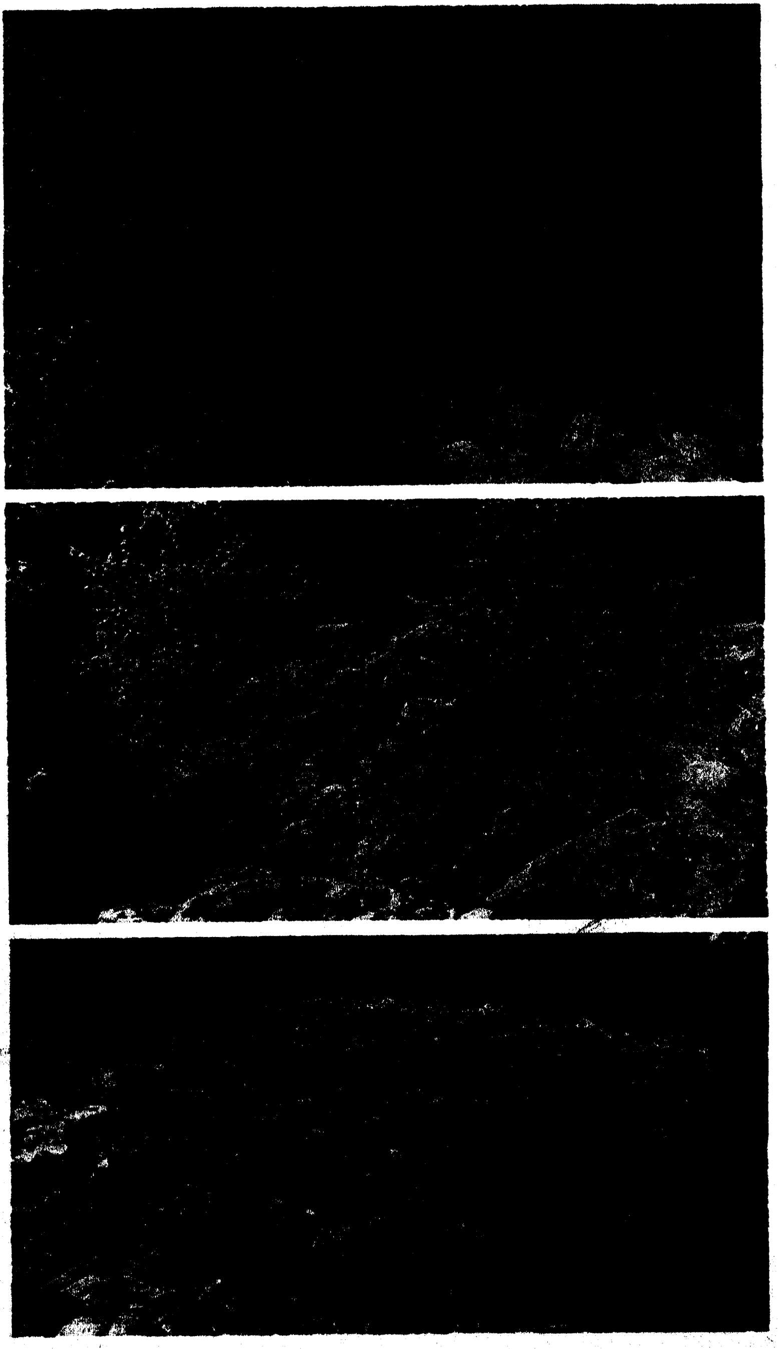
Talus and block features



Chert blocks on slopes of western Blue Mounds



Sloping block field south of Devil's Lake



Block features in Pine Creek and Baxter Hollow the talus and appeared to enter talus at a depth of 285 feet.

Indications of relative stability at present.—The following evidence points to a static condition of the talus at the present time:

1. The distribution of trees on the talus indicates that additions of new material from the weathered bluffs above have virtually ceased. Few, if any, paths for falling rock are open between trees, and, although countless trees are in vulnerable positions near the top of the talus, none were observed to show any injury from falling rock. It is concluded that falling or rolling of the rock is infrequent relative to the life-span of the trees.

2. At the top of the talus there is a striking appearance of arrested motion. The bedrock ledges are rent and riven, and blocks both large and small may be observed in all stages of detachment. Many pinnacles appear to be in a state of imminent collapse. Yet nearly all have the same discolored, weather-beaten appearance and the same mottling by lichens. Fresh joint planes are comparatively rare. All is such as to suggest an interval of vigorous breakdown of the bedrock, suddenly arrested and followed by more leisurely weathering in place.

3. The development of the linear depressions found locally at the base of the talus must be a slow process, according to Thwaites's interpretation; yet there is no indication that the process has been appreciably interrupted by the arrival of new talus blocks at the base of the slope.

4. The lichen covering and weathered appearance so typical of the talus blocks suggests an undisturbed condition for considerable time.

5. According to section foremen on the railroad along the west side of the lake, the falling of rocks on the tracks is very rare. An electric device designed to give warning of rock-falls along one stretch of track was not activated during a period of 4 years. In the spring it is customary, however, for a crew of workmen to dislodge any blocks that seem likely to fall. This undoubtedly is necessitated by the fact that the base of the talus has been artificially oversteepened along one stretch of track.

There is thus good reason for believing that growth of the talus by addition of new material is no longer in progress at an appreciable rate, although the overtopping rock ledges constitute a potential source of large quantities of material. Undoubtedly, there is some very gradual creep of the talus as a whole, but this must be slow in relation to the rate of tree growth. The basal flattening along the west shore is probably due to this factor, although the action of lake ice in winter may play a part also.

Age of the talus.—Evidence as to the age of the talus is indirect. The presence of talus at depth beneath the glacial drift indicates that talus formation was in progress before the Cary ice sheet blocked the gorge of the ancestral Wisconsin River. The unfilled depressions at the contact of the talus with drift suggest the possibility that talus building may have been essentially complete at the beginning of postglacial time. Additional evidence relating to age is provided by comparison with conditions at Baraboo Narrows (locality *A-8*), about 8 miles northeast of Devil's Lake. At that locality (pl. 1, *B*) there is virtually no accumulation of loose rock except in a few small pockets along re-entrants in the slope. Furthermore, there are no indications that the detachment of joint blocks from the bedrock is now very active. Since this locality lies inside the glacial border and was thus swept clear of pre-existing talus and loose rock, it provides a measure of the extent of postglacial weathering and block production, disregarding the small pockets of loose blocks noted above, which are so situated as to suggest remnants of older talus protected from glacial scour by their position. It may be noted that the rock at Baraboo Narrows is the same as the rock at Devil's Lake, the topography is similar, and the structure differs only in steeper dips. It therefore provides a fair basis for comparison of weathering phenomena, and reasons for the absence of postglacial block accumulation at Baraboo Narrows have direct implications for the Devil's Lake locality. Two hypotheses for the absence of talus at the former locality suggest themselves:

1. Talus building is normally so slow a process that postglacial time has not been long enough to produce appreciable results. However, if postglacial time be estimated as $\frac{1}{40}$ the duration of the Pleistocene epoch, which seems a conservative estimate, forty times the amount of loose rock now observable at Baraboo Narrows would be small indeed as compared with the volume of talus in an equivalent section of the Devil's Lake locality. Furthermore, if the rate of supply were as slow as that, it seems probable that the rate of weathering and removal on such steep slopes would keep pace with it, so that little or none of the fallen rock would be left to accumulate as talus.

2. Talus building was dependent on climatic conditions associated with glaciation and was therefore inactive during interglacial and postglacial time. On this basis the absence of talus at Baraboo Narrows and the relative stability of the talus at Devil's Lake would be due to a common cause—unfavorable postglacial climate, inhibiting the dislodgment of otherwise unweathered joint blocks in quantity. It would follow that the weathering products of nonglacial times should be distinctively different from those of glacial times, presumably characterized by more advanced stages of disintegration and decomposition. Evidence consistent with this interpretation is found in the abandoned quarries on the northeast side of Devil's Lake. There a thin veneer of talus is separated from the bedrock by about 5 feet of stony soil containing small blocks and rock fragments scattered through an earthy matrix. The contrast between the talus and the underlying material is striking and points to a marked change in conditions of weathering when talus accumulation began. Such a change is most readily accounted for as due to the coming of glaciation; and on this basis the pretalus weathering would be assigned to interglacial time and the talus building itself to Wisconsin time, or some portion thereof. If it can be shown that the processes most effective for building talus are ones which would be more favored by glacial conditions than by nonglacial conditions, the interpretation given above is confirmed. This question is considered below.

Origin of the talus.—The development of talus requires that loose rock be supplied at the top of a steep slope more rapidly than it can be removed at the bottom or be reduced by weathering on the slope. The question is mainly one of mechanical dislocation of joint blocks, at a rate which is rapid compared to the weathering of unjointed portions of the rock. The expansive force of freezing water is known to be the most powerful and rapid means of splitting rock. In modern arctic and alpine regions its effects are manifest in the general mantle of rubble wherever well-jointed hard rock occurs. Even on glacially scoured rock surfaces, in frigid regions, frost weathering has produced extensive disruption during the relatively brief span of postglacial time (Högbom, 1914, pp. 269–281; Bretz, 1935, pp. 160, 225; MacClintock and Twenhofel, 1940, pp. 1734–1737). It is thus evident that frost weathering would be adequate to account for the rapid production of talus in the Baraboo area, probably during a single glacial stage. Indeed, there is no other process to which appeal can be made for more rapid detachment of blocks than is to be observed at present. Frost weathering obviously would be more favored by glacial than by nonglacial conditions and would require, furthermore, that the glacial climate in this area be markedly colder than the present climate. A consideration of the origin of the talus thus confirms the dating of the talus arrived at from other types of evidence and carries further implications as to the nature of the climatic conditions at the time when the talus was formed. These conclusions are in essential agreement with Blackwelder's (1935) interpretation of talus in the Basin Range Province and with the concepts of Lozinski (1912) and many other European students of periglacial phenomena.

BLOCK CONCENTRATIONS AND BLOCK-STREWN SLOPES

Localities A-6 and A-7.—These localities lie on the south slope of the South Baraboo Range and are situated in the NE. ¼ of Section 28 and the NE. ¼ of Section 34, T. 11 N., R. 6 E., respectively. At both places sloping block fields of similar character have been excavated for construction material. Exposures were better at *A-7*, and for this reason it is selected for detailed description below.

The block field is irregular in outline and somewhat elongated. The surface is a bare, jumbled mass of irregular, unsorted quartzite blocks up to more than 5 feet in length (pl. 3, *A*). The blocks are mostly angular to subangular, are mottled with lichens, and have a weathered appearance. Many are split by cracks up to about an inch wide. The slope ranges from 10° to 13° and shows many minor irregularities. The block field is surrounded by wooded slopes with stony soil and has a few trees growing on its surface. The undisturbed growth of the trees, together with the weathered condition of the blocks, gives an appearance of complete immobility and breakdown in place.

The internal character of the block mass is revealed by excavations (pl. 3, *B*). The rock is entirely free of fine material to the lowest level exposed, representing a total thickness of more than 12 feet. The larger blocks tend to be segregated toward the surface. On the blocks below the surface, a heavy brown to black stain is prevalent. The percentage of interstitial open space throughout the mass is high.

Immediately to one side of the block field proper, a continuation of the excavation referred to above shows 5 feet of loose blocks similar to those described above, but with the interstices thoroughly filled with earthy material (pl. 3, *C*). The exposure is topped with a cover of sod and forest. It is evident that stripping of the soil and removal of the earthy matrix would leave a clean block accumulation like that first described.

Baxter Hollow (locality A-5).—Baxter Hollow is a youthful valley of moderate to steep slopes and a local relief of about 400 feet. It is mostly wooded. It was found to contain one block stream of fair size, several smaller block streams, and various irregular block accumulations and block-strewn slopes.

The larger block stream (pl. 4, *C*) is located on the east side of the area, in NW. ¼ of Section 33, T. 11 N., R. 6 E. It lies along a minor tributary valley and is divided into two sections by a wooded interval. The upper section is about 400 feet long, up to 75 feet wide, and has a slope of about 5°. The lower section, separated from the upper by a gap of about 250 feet, has a length of about 150 feet and a slope of about 7°. In general, the surface of the block stream is similar to those of the block fields at localities *A-6* and *A-7*. The blocks are up to 10 feet long. The surface is rough and uneven in detail. The blocks have a weathered appearance, and many are cut by open cracks. The walls of some of the cracks show a degree of weathering similar to that on the block surfaces, indicating that the rate of present-day creep is very slow as compared with the rate of weathering. The block stream is entirely surrounded by trees and has a few trees on its surface.

Several minor block streams are located on the east side of the road, along the western edge of the SW. ¼ of Section 28. They occupy slight indentations in the slope and range up to about 10 feet in width and up to several tens of feet in length. The lower ends of some are exposed in a road cut, where it is seen that their thickness is only a few feet. They are underlain by an earthy rubble zone containing a high percentage of stony material to a minimum depth of 6 feet. Removal of the fine material from this rubble by running water would leave a residual accumulation of blocks like those of the block streams.

At other places in Baxter Hollow, particularly in the SE. ¼ of Section 29 and in the NW. ¼ of Section 33, the occurrence of large blocks embedded in the soil is common, and locally there are discontinuous accumulations of blocks, some of which are associated with outcropping ledges. The impression one receives is that stripping of the surface soil at most places would reveal the subsoil to be a mass of blocky rubble.

Locality A-3.—This locality lies somewhat less than a mile northwest of Devil's Lake (NE. ¼ Sec. 14, T. 11 N., R. 6 E.). Scattered blocks and boulders of quartzite, sandstone, and conglomerate occur along a shallow valley and adjoining gentle slopes. Some of the blocks are almost buried in the soil, while others appear to be largely above the ground surface. Locally the blocks are jumbled together.

Age and origin.—Except for slope, the surface appearance of the block fields described above is similar to that of the talus at Devil's Lake and suggests that their age may be the same. Beyond this surmise, the question of age is closely linked to the question of mode of accumulation.

In considering the latter, the evidence against purely residual accumulation may be reviewed first: (1) the same degree of weathering is found from top to bottom; (2) the segregation of the larger blocks toward the surface is exactly the opposite of what would be expected in a residual weathering deposit; and (3) the marked discordance in shape, size, and orientation between individual blocks, together with the large proportion of open space, is entirely incompatible with the idea of mere modification and loosening of rock in place by weathering along joint planes. Some degree of transportation is indicated, although quantitative data on the amount of movement are lacking.

The material of the block fields is essentially similar to that in the talus discussed in preceding pages and is best explained as having been supplied in the same way, by vigorous mechanical weathering through frost action. The mode of transportation, however, must have been very different from that responsible for talus building—some process capable of working effectively on low to moderate slopes.

In considering the question of transportation, it may be asked, first, whether movement of the constituent parts of the block fields could have taken place in their present condition. Certainly, the internal friction of a mass of loose rock without interstitial fine material would seem prohibitively high. The feasibility of movement would seem much greater if a matrix of earth, or possibly of ice, were present to provide lubrication, prevent tight contact between blocks, and transmit pressure. The occurrence of an earthy matrix in the deposit immediately adjoining the block field at locality *A-7* and in the rubble under the minor block streams in Baxter Hollow strongly suggests the possibility that such a matrix was originally present throughout, later to be removed locally by surface and subsurface waters. The conspicuous staining of the buried blocks in the block fields also is best explained on that basis, with the fine material serving to keep mineralized waters in contact with the blocks much longer than if the interstices were open. The possibility of an ice matrix is suggested by Capps's studies on rock glaciers in Alaska (1910), but Capps provides no precedent for attributing any importance to interstitial ice on slopes of less than 9°; hence this factor seems admissible only in an auxiliary role. It may be concluded that the block fields and block streams represent local modifications of a more widespread mantle of rubble, which is composed of material supplied by mechanical weathering and transported by some process of mass movement.

At present there is no indication that either mechanical weathering of bedrock or mass movement is proceeding at a significant rate. The weathered appearance of the blocks at the surface, the cracking of blocks without appreciable separation of the broken pieces, the undisturbed growth of trees on and around the block fields, and the heavy, unmarred staining of buried blocks all point to present stability. The probability that the bare, blocky tracts were produced by removal of interstitial fines is in keeping with this interpretation. The fact that the above conditions are characteristic of both the steeper and the more gently sloping block accumulations indicates that immobility is not merely a matter of having attained a "profile of equilibrium." It can be concluded only that the various blocky accumulations antedate the present and could have been formed only at a time when climatic conditions were sufficiently different to promote the action of processes which are no longer effective. The only type of climatic shift which could have been favorable for the required processes would have been one to lowered temperature, such as might be expected to have accompanied glacial stages. Such a change of climate would have caused intensified frost weathering to supply the rubble and would have inaugurated solifluction and frost heave to move the rubble downslope. This explanation was applied to the "stone rivers" of the Falkland Islands by Andersson, in 1906, on the basis of comparison with modern arctic phenomena. A similar explanation, though less clearly formulated, had previously been applied to rubbly deposits in England (Geikie, 1894). Andersson's interpretation was later extended to block fields and rubbly accumulations in many parts of Europe by numerous workers (Salamon, 1917; Meyer-Harrassowitz, 1918; Kessler, 1925; Quiring, 1928; Fourmarier, 1933; Büdel, 1937; and Dines *et al.*, 1940). Büdel's work was of outstanding significance. He found that coarse rubble had moved as far as 2 km., and on slopes as low as 2°. Positive evidence that the accumulation of this material had ended with the waning of the glacier and that stability had prevailed throughout postglacial time was provided by the fact that the rubble was overlain by undisturbed bog deposits containing pollen sequences representing the entire postglacial interval and beginning with a subarctic birch-pine flora. Postglacial stability was found to be characteristic of block accumulations having slopes up to more than 17°. Although the above type of evidence has not yet been found in the Baraboo area, Büdel's results do provide a basis for comparison which strongly favors the interpretation set forth above for the Baraboo area.

It may be concluded that the block fields are a result of processes peculiar to periglacial conditions and that the primary accumulation of rubble dates back to Wisconsin time, or possibly in part to pre-Wisconsin glacial conditions. Like the talus, the block accumulations required intense frost weathering; unlike the talus, they required, in addition, special processes of transportation, also related to frost action. The fact that these same processes of transportation did not effect a more complete removal of the talus material itself may be attributed to one or more of the following conditions: (1) more rapid weathering because of more exposed position of parent ledges, and hence more rapid supply of blocks; (2) virtual absence of interstitial fine material in the talus, so as to make for greater frictional resistance to movement; (3) better drainage of the talus due to steeper slope.

CHOKED VALLEYS AND BLOCK CASCADES

Description.—Choked valleys and block cascades are well developed at localities *A-1* (NE. ¼ Sec. 28, T. 11 N., R. 5 E.) and *A-2* (NW. ¼ Sec. 17, T. 11 N., R. 6 E.). At the latter locality a block-strewn valley floor leads downstream to a block cascade, which marks the edge of a benched portion of the valley bottom. The block cascade (pl. 4, *B*) drops steeply some 10 feet. The blocks are large and heterogeneous and have the appearance of being jammed together, presenting a rough and tumultuous aspect. Running water can be heard beneath the blocks. Trees and bushes are growing among them. Jumbled masses of blocks border the cascade on both sides. A well-weathered appearance is general.

On the upvalley side of the block cascade, some blocks are scattered at random, others are clustered, and still others occur in elongated strips. At some places no surface stream is recognizable as such. The sound of running water can be heard locally, however, under the blocks in the elongated strips (pl. 4, *A*). Where these strips end, the course of the underground steamlet may be traced, in part, along a series of holes in the sod, through which the sound of running water comes. The drainage, at least during times of ordinary flow, is entirely under the surface at these places and appears to be in process of undermining the sod, exposing more blocks and extending the blocky strips by washing away the interstitial fine material from a coarse, rubbly deposit.

At locality *A-1*, a shallow, rounded valley bottom funnels into a miniature gorge with steep bluffs of quartzite. Above the gorge the valley is littered with blocks, which are most conspicuous along the shallow, grassy depression which constitutes the stream channel. Toward the gorge the blocks become more numerous. On entering the gorge, the valley narrows, and its gradient steepens, first gradually to about 2° and then more abruptly to about 8°, in descending over a cascade of huge blocks. Below the cascade the gradient decreases to about 5°, and the valley gradually widens; a band of coarse, jumbled blocks persists along its center for some distance.

Interpretation.—The condition of the valleys described above is aberrant in terms of normal erosional development. The streams are powerless to move the coarse blocks over and around which they flow and will remain so until these blocks have been reduced to much smaller particles by weathering. Obviously, these streams could never have formed the existing valleys if coarse blocky detritus had been continuously supplied at such a rate as to produce the condition now observed. The influx of blocks may therefore be regarded as an abnormal interruption of the normal processes of stream erosion, literally choking the valley bottoms with coarse debris. At the present time, normal stream action appears to be in process of becoming re-established but is engaged merely in the removal of interstitial fine material from the rubbly fill. If carried far enough, this may convert the valley floor into a barren block field.

The choking of valleys with coarse detritus represents simply a further and particularly striking step in the general process of mass movement already invoked to account for the rubble and the block concentrations elsewhere in the Baraboo area. The fact that it can be seen here to have grossly altered the normal progress of erosion serves to direct attention to its disharmonic character with respect to the cycle of stream erosion and to emphasize the fact that the general process could have taken place only under conditions notably different from those of today. Solifluction and frost heaving of detritus supplied by accelerated mechanical weathering under periglacial climatic conditions during one or more glacial stages provide an adequate explanation.

The block cascades are most satisfactorily explained as representing a local variation of the general process of mass movement. They bear some resemblance to the boulder-banked terraces of arctic and alpine regions (Eakin, 1916); the latter are known to be associated with vigorous frost action, but their exact mode of formation is not entirely clear. Possibly the block cascades are roughly analogous to ice jams or log jams in rivers and were related to a down-valley component of the mass movement.

THE BLUE MOUND AREA

Description.—The location of the Blue Mound area is shown in figure 1, and the generalized topography in figure 3. The "mound" is an erosion residual, standing some 500 feet higher than the surrounding country. It is held up by the Niagara chert (Silurian), which is underlain by shale (Martin, 1932, pp. 66-67). Owing to inadequate exposures, the thickness of the chert and the extent to which limestone may be associated with it are uncertain. The topography, however, suggests that thickness is not greater than 100 feet. The slopes of the mound are largely wooded, with scattered clearings. The Wisconsin drift border lies approximately 10 miles to the east.

The feature of particular interest on Blue Mound is the distribution of chert blocks around the slopes. The blocks are found at distances of up to nearly 1 mile from the summit and as far as 600 feet below the summit. Particularly prominent are the large scattered blocks on the east slope (pl. 1, *C*) and the block concentrations on the north slope (pl. 2, *B*).

The eastern slopes of Blue Mound are gentle and broadly undulatory; chert blocks are widely distributed and are

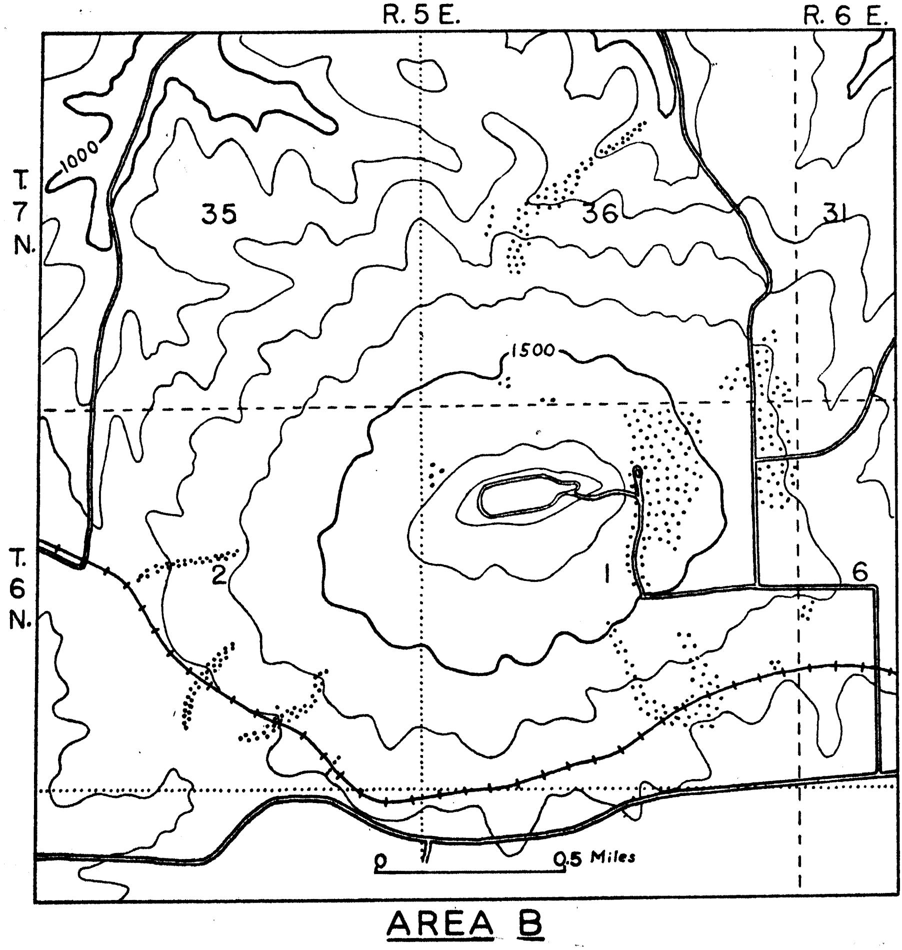


FIG. 3.—Generalized map of the Blue Mound area. Contour interval is 100 feet, and contours are based on the Blue Mound topographic sheet. Roads are drawn from aerial photos of the U.S. Department of Agriculture. Block-strewn areas are shown by stippling and are approximately complete only for cleared areas.

particularly conspicuous between the 1,500- and 1,600-foot contours. The proportion of very large blocks is high, with many ranging from 10 to 25 feet in length and some standing as high as 7 feet above the ground surface. The individual blocks are very irregular in form. They appear to be stranded where they lie and to be undergoing gradual breakdown by weathering (pl. 2, *A*). Irregular cracking of the rock is common, and detached fragments are seen on and around many of the blocks. Many of the blocks occur on slopes of between 3° and 4°. Individuals are mostly well separated, with virtually no tendency toward clustering or concentration.

On the north side of Blue Mound an irregular block field passes downslope into a narrow block stream which follows a small valley for some distance. In the area shown in plate 2, *B*, the average slope is about 7°. The blocks are similar in form and general appearance to those on the east slope, but the average size is smaller; a majority probably range between 5 and 12 feet in length, and few reach 20 feet. Although the individual blocks are crowded fairly close together, most of them are separated by sod. Evidence of present stability is similar to that for the east slope, but with the significant addition of one perched block (pl. 2, *C*), which obviously is in too precarious a position to stay perched very long if appreciable movement were now in progress. Except for a small, grassy channel on one side of the blocky area, drainage apparently is by sheet wash, and no appreciable erosion by running water seems to be taking place in this section.

Downstream, the block field becomes a block stream following a small valley. It narrows to a mere ribbon as the valley deepens and steepens its gradient. At the lower end, stream action is actively denuding the blocks by removal of the soil in which they are embedded. This process appears to be working headward.

On the southwestern side of Blue Mound, small, elongate blocky areas occur in shallow valleys. Locally, where gullying has stripped off the soil, the volume of blocks in the subsoil is seen to be considerably greater than might be expected from surface indications.

Undoubtedly, many other spreads and concentrations of blocks are hidden in the wooded portions of the slopes on Blue Mound. The occurrences described above, however, are believed to be representative and to provide an adequate basis for interpretation.

Interpretation.—Two possible explanations may be considered for the distribution of the blocks around Blue Mound: (1) as a lag deposit of blocks let down over steep slopes by creep and sapping to successive positions of the base of the escarpment during its gradual recession to the present stand and (2) as a result of transportation over existing slopes from approximately the present position of the parent ledges, by processes no longer active. The first of the above hypotheses is essentially the one suggested by Davison (1889) for the origin of the stone-rivers of the Falkland Islands, which was never generally accepted. As applied to the Blue Mound area, this hypothesis would require the accumulation of the blocks to have extended over a span of time sufficiently long for the escarpment to have retreated more than ½ mile to its present position. Since the erosional retreat of such an escarpment is a very gradual process, this would imply that the outermost block accumulations are of much greater age than those closest to the present rim of the escarpment, probably dating back to middle or early Pleistocene time. If that were the case, there should be a marked difference in the size and state of weathering of blocks at different distances from the present hill. This supposition, however, is not in accord with the observed facts, for all the blocks show essentially the same degree of weathering, as nearly as can be ascertained, and size distribution appears to be random on any given sector of the slope. A further difficulty with the above hypothesis is that, even if it could account for the block spread on the eastern slope of the mound, it would still fail to account for the concentration of blocks in valleys, where valleys occur, and for the evident choking of some of the valleys. If the hypothesis of lag accumulation were to apply to such occurrences, it would have to be modified to admit secondary concentration by supplemental processes, and any such processes which could be admitted would in themselves be adequate to account for the observed features.

According to the second of the above hypotheses, the present distribution of the blocks would be explained as due to essentially the same processes, working under the same climatic conditions, as those discussed in connection with the block accumulations of the Baraboo area. The evidences supporting this interpretation are essentially similar to those in the Baraboo area and need not be recounted, but only supplemented by the following statements:

1. The geological conditions at Blue Mound are particularly favorable for solifluction, owing to the shale underlying the block-forming rock and providing a source of mud.

2. The stratigraphic and topographic localization of the block-forming rock at Blue Mound makes possible a definite estimate of the distance which the blocks have traveled and thus lead to confirmation of deductions as to the importance of downslope movement in the Baraboo area.

3. It is possible that detachment of the blocks at Blue Mound may have been effected as much by undermining and collapse, through flowage of the underlying shale, as to frost weathering. In fact, the observed irregularity in the form of many of the blocks is hardly suggestive of joint control unless it be assumed that the original joint blocks were greatly modified by subsequent weathering.

4. The perched block in plate 2, *C*, suggests the possibility that it was left in that position by solifluction, later to become "emergent" as the enveloping mud flowed on or was washed away; an alternate explanation, however, is that it was separated from the adjoining blocks by differential weathering. The uniqueness of the occurrence, however, is unfavorable to the latter interpretation.

5. It is possible that the largest of the blocks, particularly on the eastern slope, represent unit movement rather than mass movement. This would imply that they moved as miniature landslides, after the fashion of the masses which I described from Canjilon Divide, New Mexico (Smith, 1936). In that case the process would have involved seasonal frost heave normal to the ground surface, followed by gravitational settling during the time of thaw, either in a vertical direction or in a steep downslope direction.

PROBABLE PERIGLACIAL FEATURES AT OTHER PLACES

In the course of reconnaissance study and searching of the literature, it was found that features consistent with the hypothesis of periglacial origin, although not necessarily accompanied by definitive evidence thereof, occur also at the following places.

Locality C.—This locality is situated on the west side of Wisconsin Highway 39, about 0.3 mile northwest of the town of Mineral Point (N. ½ Sec. 36, T. 5 N., R. 2 E.). Sandstone outcrops at the top of the slope, and sandstone blocks litter the surface below. Toward the top they form a talus, with a slope angle of 36°; this passes downward into a small, irregular block field with a gradient down to 6°. The blocks range up to 15 feet in length. Many show appreciable rounding of the corners and edges by weathering. Locally the blocks appear to override one another, giving a suggestion of rude imbrication. A majority of the blocks is more or less surrounded by sod. The appearance is one of complete immobility. There is no apparent reason for doubting that the blocks accumulated here in the same general way as did those in the block fields of the Baraboo and the Blue Mound areas.

Locality D.—This locality lies about 4 miles southeast of Dodgeville, along a branch of the Illinois Central Railroad (near the western edge of SW. ½ Sec. 7, T. 5 N., R. 4 E.). Attention was directed to this locality by Sardeson's (1897) description of deposits supposedly made by local glaciers. The exposures noted by Sardeson were found to be entirely overgrown at the time of my visit, but his description suggests a solifluction deposit, and the occurrence of large blocks of rock littering gentle slopes bordering the valley near by is consistent with this explanation.

Locality E.—This locality lies on the north side of the creek at Jonesdale (near SW. cor. Sec. 16, T. 5 N., R. 4 E.) and was mentioned by Sardeson. A heap of sandstone blocks, now largely overgrown by forest, lies at the foot of a steep bluff. It appears to be stable at present and could be explained satisfactorily as a relict feature formed under periglacial conditions.

Locality F.—This locality is a road cut along U.S. Highway 12, 2.2 miles east of the town of Tomah. Numerous well-developed ventifacts of hard, siliceous rock, up to 8 inches long, are scattered through a structureless, sandy soil to a depth of about 2 feet. The ventifact zone is underlain by sandstone and is covered with grass and trees. No drifting sand is found anywhere in the vicinity at the present time. The association of sandblasting with the periglacial environment is well known, and it seems reasonable to suppose that the ventifacts here may have originated under periglacial conditions. The vertical distribution of the ventifacts might then be attributed to periglacial frost churning, as noted by Bryan (1932) for localities in southeastern Massachusetts. Obviously, further study would be required to test this hypothesis.

Alluvial terraces of the Kickapoo Valley.—Alluvial terraces along the Kickapoo Valley have been described and interpreted by Bates (1939). The alluvial fill is said to be composed almost entirely of material brought down from the valley sides by slope wash and creep, without appreciable re-working by the axial stream. It is suggested by Bates's discussion, though not explicitly stated by him, that aggradation was due to a quickening in the rate of supply of material from the valley sides. In discussing the origin of the terrace, however, after making correlations with the terraces of the Wisconsin River, Bates infers that fluvioglacial aggradation of the Wisconsin

River was an adequate cause for the aggradation of the Kickapoo. Just what effect the aggradation of the Wisconsin River could have on slope wash and creep along the Kickapoo is not explained. The possibility of climatic influences on erosion and deposition, working entirely within the Kickapoo drainage basin (Martin, 1932, p. 133), is completely ignored. To me it seems reasonable to postulate that the aggradation of the Kickapoo Valley was contemporaneous with the choking of valleys in the Baraboo area and was brought about in much the same way—by the influence of perglacial climate on slope processes. Provisional interpretation of the terraces as another type of periglacial phenomenon thus seems justified.

DISCUSSION OF PERIGLACIAL PROCESSES

In this section it remains to particularize somewhat further as to the working of the geomorphic processes referred to on preceding pages. The discussion is based to a large extent on observations of present-day conditions in arctic and subarctic regions, as recorded by Andersson (1906), Högbom (1914), Eakin (1916), Bretz (1935), Taber (1943), Muller (1947), Washburn (1947), and others. The results of studies in the former periglacial zone of Europe, as set forth by Kessler (1925), Büdel (1937), Poser (1947), and other workers, are drawn upon also.

The periglacial processes of particular concern are mainly mechanical weathering and mass movement on slopes. Mechanical weathering by frost wedging was important in providing a continued supply of broken rock. Owing to lower temperature, frost penetrated more deeply and exerted greater force. Chemical weathering, on the other hand, was retarded. With the onset of periglacial conditions, frost action began its work on a weathered zone inherited from interglacial time and developed in part by chemical weathering. Its first effects probably were to break down unsound rock fragments and, by working along the many joint cracks and other openings already prepared by chemical weathering, to produce a relatively fine rubble. As this was gradually removed by gravity and by frost-induced slope processes, the frost came to act on deeper, fresher zones of bedrock, in which there had been less preparation by previous weathering. Fewer potential joints had been opened, and probably frost alone was relatively incompetent to open them. As a result, only the more persistent, throughgoing joints offered points of attack; and, as these were more widely spaced than the minor joints, the effect was to make for the dislocation of larger and larger joint blocks as frost worked to deeper levels. It is believed that this, in conjunction with mass movement, provides one possible explanation for the segregation of the larger blocks in the surface zone of at least some of the block fields.

In the paragraph above, a massive rock, relatively impervious except along joints and widely spaced bedding planes, is assumed. In rocks having more closely spaced openings the effects would naturally be different, and the particles loosened would be much smaller. In the case of a shale, the resulting product would be mud.

The removal of loose rock provided by mechanical weathering was effected directly by gravity where the slope exceeded the angle of repose, forming talus. Where the angle of slope was lower, a more gradual process of creeping was operative. The loose rock, intermixed with pre-existing soil and with such additional fine material as was supplied by periglacial weathering, was slowly carried downslope to form rubble sheets or block-strewn surfaces.

The downslope movement of detritus under periglacial conditions is believed comparable to the mass movement of saturated material observed today in arctic regions, and generally designated by the loosely defined term "solifluction." This process is reported to be far more rapid and effective than the ordinary variety of hillside creep at work in temperate regions. The movement occurs in the season of thawing and is at least partly by viscous flowage. Saturation is effected by water from melting of snow and/or interstitial ice. The melting of interstitial ice and of the actual segregations of ground-ice typical of frigid regions is particularly important. Where the volume of ground-ice is large, the volume of water liberated may be excessive, leading to what may be termed a "state of supersaturation." The effect of the water thus liberated is to add to the weight of the soil mass, to act as a lubricant, and, where it cannot escape readily, to carry part of the load, leading to fluid-like behavior.

Maintenance of saturation is important for the process of solifluction and requires the prevention of loss of water by downward seepage. The presence of a perennially frozen subsoil is particularly effective in this connection and is believed to be a factor of primary importance for solifluction in arctic regions. Positive evidence for the former prevalence of perennially frozen ground in the Driftless Area has not yet been found, but it is not unreasonable to infer that such a condition did exist, and the process of mass movement is easier to explain on that basis. In the former periglacial zone of Europe, definite evidence for once perennially frozen ground has been reported, thus providing a precedent.

The above discussion of solifluction applies particularly to fine-grained material, which can be converted into a muddy mass. Such masses, however, have been observed to carry along a considerable amount of much coarser material with them. As to the maximum proportion of coarse material that may be present without inhibiting the process of flowage, observational data are simply lacking. In the Blue Mound area an ample supply of mud probably could have been derived from the shale underlying the chert, thus to maintain a suitable ratio of fine to coarse material; hence the above question is not pressing. In the Baraboo area, however, the proportion of coarse material observed in the subsoil at localities such as *A-5* and *A-7* is so high that interference with simple flowage would seem probable. Some modification of simple soil flow would seem required here, and for this there is little direct precedent in observation. One possibility is that the observed blocky accumulation represents a lag deposit gradually banked up from the passage of a much larger volume of earthy material, in a manner more or less analogous to the development of a log jam in a river. It is somewhat difficult, however, to point to an adequate source for a much larger volume of fine material than is now present. Another possibility, and one which seems more acceptable, is that downslope movement was effected partly by annual alternations between surface expansion or upward heaving by frost and downslope settling upon thawing. This certainly seems reasonable for individual blocks and conceivably could apply to the rubbly mass more or less as a whole. The amount of annual movement would not need to be great—0.1 foot per year would amount to 1,000 feet in 10,000 years. Until further critical studies are forthcoming from regions where mass movement is now active under frigid conditions, however, this question must remain speculative.

However the movement took place, its effects were to spread a mantle of rubble over the slopes, filling up low places, choking minor valleys, and disrupting normal stream action. These effects probably were cumulative through more than one substage of Wisconsin time, and it seems not impossible that a pre-Wisconsin glacial stage also may have contributed to the observed effects.

When deglaciation brought climatic amelioration, frost weathering lost its effectiveness, and mass movement came virtually to a halt. The immobilized rubble sheets thereafter were subject to breakdown and decay by the very gradual action of normal weathering processes. Locally, the fine material between the blocks was washed away to produce the barren blocky surfaces so striking in the present landscape.

CONCLUSIONS

1. During at least a part of Wisconsin time, the climate of the Driftless Area was more rigorous than at present and had some similarity to that now founl in subarctic regions.

2. As a result of periglacial climatic conditions, geomorphic processes differing both in degree and in kind from those now observable were at work in the area. Slopes were blanketed with detritus supplied by accelerated mechanical weathering and carried downward by invigorated mass movement and related processes.

3. Post-Wisconsin erosion has made comparatively slight progress in effacing the results of periglacial processes. Many of the detailed features of the landscape date back to Wisconsin time.

4. At many places a periglacial rubble zone rather than a residual weathering zone constitutes the parent material for present-day soils.

5. Much remains to be learned about the types and distribution of periglacial phenomena in the Driftless Area and their significance in the interpretation of climatic and geomorphic history.

Acknowledgments.—Particular thanks are due to Professor J Harlen Bretz and to Professor F. T. Thwaites for suggesting localities to be investigated, and to the latter also for stimulating discussions both in the field and in the office. Acknowledgment is due also to Dr. A. D. Hasler for the facilities provided by the Geneva Lake Institute of Natural Science. Field expenses were defrayed by Grant No. 391-41 from the Penrose Fund of the Geological Society of America. The paper is published here with the permission of that Society.

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THE FROST-MOVED RUBBLES OF JUMBO DOME AND THEIR SIGNIFICANCE IN THE PLEISTOCENE CHRONOLOGY OF ALASKA[1]

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ABSTRACT

Jumbo Dome, a prominent landmark on the north side of the Alaska Range, is a small body of intrusive andesite surrounded by schist and by poorly consolidated sediments of Tertiary age. Frost-moved rubbles, consisting of coarse andesite blocks, almost completely mantle the dome and have advanced outward from it across a gently sloping terrain for distances as much as 1½ miles. Several different periods of rubble development are recognized, based on the amount of vegetal covering of the deposits, the preservation of their surface forms, and their degree of erosion by fluvial processes. The rubbles are not now moving and are believed to have originated under the influence of an arctic climate in a manner analogous to rock glaciers. The difference in altitude between presently moving rock glaciers in this region and the rubble deposits of Jumbo Dome corresponds to the difference in altitude between present ice-filled cirques and the lowest cirques of the Wisconsin stage of glaciation. Reasons are given for believing that fluvial weathering and destruction of rock glaciers represent climates at least as mild as the present. On this basis five separate glacial episodes, separated by interglacial and interstadial epochs, are recognized.

INTRODUCTION

A unique assemblage of detrital deposits surrounding Jumbo Dome, a prominent landmark on the north side of the Alaska Range, offers the possibility of establishing a more complete Pleistocene chronology for Alaska than has been recognized heretofore. These detrital deposits are believed to have originated under the influence of arctic or glacial climatic conditions in a manner analogous to that of rock glaciers. The deposits were noted during the mapping of the Nenana coal field, a part of the Geological Survey program of mapping the coal resources of Alaska. Although the field party was in the neighborhood of Jumbo Dome for six weeks, only four days in August, 1947, were spent on and about the dome. The primary aim of the work was the determination of the stratigraphy and structure of the Tertiary rocks; consequently, the features described in this article were observed only in passing, and this paper reports a reconnaissance rather than a detailed investigation. In 1948 the area was revisited for three days, one day of which was spent in the vicinity of Jumbo Dome. The detrital deposits, along with other geology, were plotted on trimetrogon aerial photographs taken by the United States Army Air Forces. Fortunately, the line of flight passed directly over Jumbo Dome, so that most of the contacts of critical importance were plotted on vertical photographs. The information plotted on the photographs was transferred to a topographic map of the Nenana coal field (U.S. General Land Office, 1916; scale 1:31,680, contour interval 50 feet). This map with slight alterations is the base for figures 2 and 3.

Jumbo Dome is a prominent steep-sided butte located in the foothill belt of the north flank of the Alaska Range, about 60 miles S. 30° W. from Fairbanks, Alaska (fig. 1). It rises to an altitude of 4,505 feet, 1,500–2,500 feet above the surrounding low, rolling country. It has a summit plateau covering an area of

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[1] Published by permission of the Director, U.S. Geological Survey. Manuscript received December 14, 1948.

about 60 acres, above which rise a few gentle hills and rock ledges, 50–150 feet high.

The northern foothill belt of the Alaska Range is 20–25 miles wide and consists of subparallel east-trending ridges and valleys, crossed by a superposed north-flowing drainage emerging from the Alaska Range. On leaving the foothill belt, the streams flow for distances of 20–50 miles across an alluvial lowland called the Tanana Flats, before joining the Tanana River, a major tributary of the Yukon.

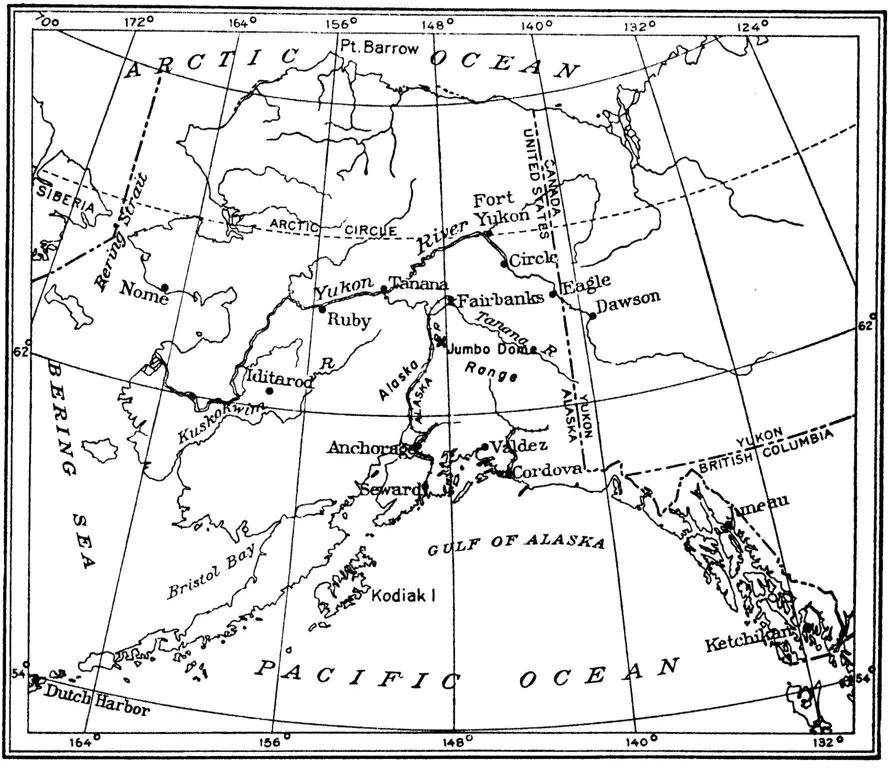
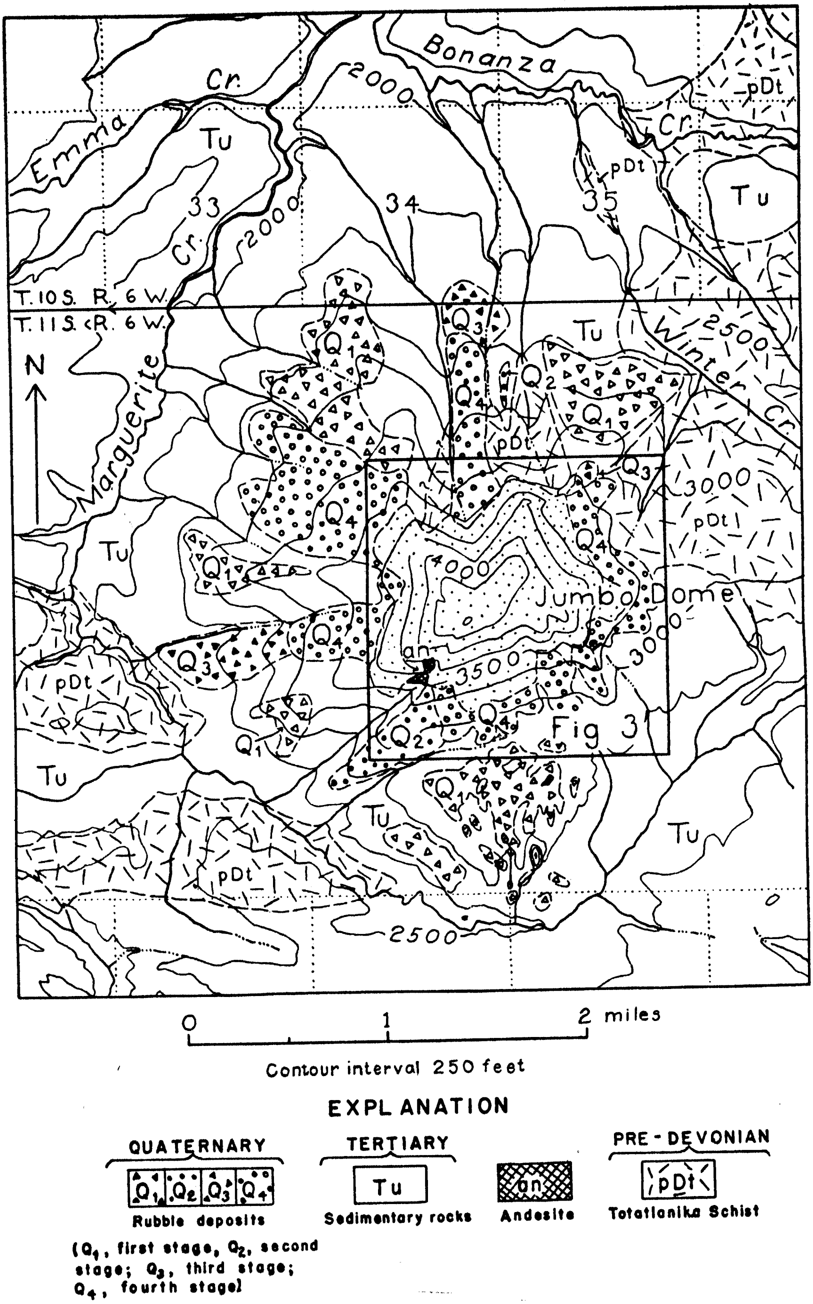


FIG. 1.—Index map of Alaska, showing location of Jumbo Dome

GENERAL GEOLOGY

The ridges of the foothill belt on the north flank of the Alaska Range are underlain for the most part by crystalline schists of Paleozoic and pre-Cambrian age and by intrusive rocks. A blanket of consolidated deposits of early Tertiary age once covered this foothill belt but is now preserved largely in the lowlands. However, the Nenana gravel—a conglomerate of middle Tertiary age—forms ridges and plateaus that rival in height some of the schist ridges. The regional geology has been described by Capps (1912; 1932, pp. 219–300; 1940, pp. 93–133).

Jumbo Dome is an intrusive mass of andesite, about 1 mile long in an east-west direction (fig. 2) and about $\frac{3}{4}$ mile wide. Platy and linear flow structure are well developed in the andesite and indi-



For geology of stippled area, see fig. 3

FIG. 2.—Geologic map of Jumbo Dome and vicinity, Alaska

cate that its intrusive contact with the crystalline schists dips steeply inward beneath the andesite, at angles of more than 45°. The mass of schists on the north and east flanks of the mountain is thought by Capps (1912, p. 25) to be Paleozoic. If so, then the intrusion is Paleozoic or younger; it is quite definitely pre-Tertiary.

Tertiary coal-bearing deposits, consisting of poorly consolidated sandstone and shale, with interbedded sub-bituminous coal beds as much as 30 feet thick, border the andesite on the west and south. As exposed in gullies on the southeast and southwest sides of the mountain, the contact between coal-bearing rocks and andesite dips 70° N. It is apparently a fault contact, for the coal-bearing formation is not baked or altered along the contact and the rank of the coal appears unchanged. The coal-bearing rocks are folded into a tight anticline immediately adjacent to the fault, but they flatten to the south, and less than a mile from the south side of the mountain they are flat-lying. On the west and north sides of the mountain, where the Tertiary rocks dip away from the mountain at low angles, the contact between the andesite and Tertiary rocks was not seen. It is presumed that the Tertiary rocks were deposited unconformably on the andesite, as similar and related bodies of igneous rocks farther east are clearly overlain unconformably by coal-bearing rocks of Tertiary age. However, no pebbles of the andesite of Jumbo Dome have been found in the coal-bearing rocks.

Following the deposition of the Tertiary sediments, the region was folded into gentle east-trending anticlines and synclines. Deformation of the rocks underlying the Tertiary deposits was by shearing along pre-existing planes of weakness. The andesite of Jumbo Dome apparently was forced upward as a roughly cylindrical-shaped mass to a vertical height of 1,500 feet above the base of the surrounding Tertiary rocks. The movement appears to have been along a circular fault zone that in part coincided closely with the steeply dipping intrusive contact of the andesite and in part was the upward continuation of this roughly cylindrical surface into the overlying Tertiary rocks. The uncertainties regarding the structure of the area do not affect the problems here considered. The only important point is that the nearly equidimensional mass of andesite is separated from the surrounding rocks by nearly vertical contacts.

GLACIATION

Jumbo Dome lies 10–15 miles north of the northern limit of Wisconsin glaciation in the Alaska Range. A terminal moraine in the valley of the Nenana River, tentatively identified as Wisconsin in age (Capps, 1932, pp. 289–291), is found about 12 miles southwest of Jumbo Dome, near the town of Healy, at an altitude of 1,650 feet. The lower limit of glacier accumulation of the Wisconsin stage, as indicated by the lowest cirques, differs considerably from place to place within the Alaska Range. In the vicinity of Mt. McKinley, 95 miles to the southwest of Jumbo Dome, it is about 4,500–5,000 feet. On the south side of the range it is even lower. It appears to rise sharply northward across the range, because of the falling-off of precipitation (Capps, 1940, p. 154), and in the vicinity of Jumbo Dome it is doubtful whether mountains lower than 6,000 feet supported glaciers at any time during Wisconsin time. At present the lowest mountains on the north side of the Alaska Range which have cirques that contain glaciers are about 7,200 feet in altitude. Thus the lowering of the orographic snowline was about 1,200 feet in Wisconsin time and probably more during earlier glaciations. Jumbo Dome (4,505 feet) lay below the lower limit of accumulation of ice in Wisconsin time.

Morainal deposits and erratics of a stage of glaciation earlier and of greater extent than the Wisconsin have been recognized in this area (Capps, 1931, p. 7). The earlier glaciers apparently advanced northward across a now almost completely destroyed surface having a present altitude of about 4,000 feet in the vicinity of Jumbo Dome (Wahrhaftig, 1946, pp. 134–135). All the rubble deposits described in this paper are found on topography which has been cut into this surface and are therefore younger than the earlier stage of glaciation.

DESCRIPTION OF THE RUBBLE DEPOSITS

The rubble deposits which surround Jumbo Dome consist of blocks of andesite that originated on the steep slopes of Jumbo Dome and extend outward across the surrounding schists and Tertiary sedimentary rocks as far as 1¼ miles from the base of the dome. Some of the masses of rubble moved outward across an eroded surface on slopes much less than 15°. Rubble deposits of several ages can be distinguished by their relative degrees of weathering and dissection. In all, five ages of rubble have been recognized. It is likely that more ages could be recognized, but adequate criteria for their distinction were not found. As will be brought out more fully later, in a section dealing with the influence of climate on denudation in the Alaska Range, it is the thesis of this paper that each rubble deposit near Jumbo Dome was formed under the arctic climate of a glacial stage or substage and then later weathered and dissected during a milder interglacial or interstadial episode.

The rubble deposits of Jumbo Dome resemble rock glaciers in many respects and probably originated in a somewhat similar fashion. The term "rock glacier" was first applied by Capps (1910, p. 360) to tonguelike masses of coarse debris in the Wrangell Mountains of Alaska, which he believed moved with a glacierlike motion. Deposits resembling these in part were called "rock streams" by Howe (1909) and "moving talus" by Bretz (1935, p. 192). Rock glaciers which have been described in the literature are large, commonly tonguelike masses of very coarse angular debris which extend down gentle slopes from the base of the cliffs or talus slopes from which their constituent fragments were derived. They are commonly 20–200 feet thick, the thickness being greatest at the lower end. Typical modern and recent rock glaciers, such as those described by Capps, Howe, Chaix (1923, pp. 1–38), and Kesseli (1941, pp.203–227), have steep fronts and sides and bear on their upper surface a series of parallel looped ridges, which are also parallel to the sides and front of the rock glacier.

The rock glaciers heretofore described are characteristically valley features, forming in cirques, and do not commonly originate on smooth slopes or form as continuous sheets, as are shown in figure 5, *A*. However, other rubble deposits of Jumbo Dome, which lie in a canyon on the north side of the mountain (see fig. 2) resemble in every respect rock glaciers described by Capps. In some respects the rubbles of Jumbo Dome have many features in common with the solifluction sheets and lobes recently described by A. L. Washburn (1947, pp. 86–96, pls. 23, 24) from Victoria Island.

It is not the major purpose of this paper to discuss whether or not the rubbles of Jumbo Dome should be considered rock glaciers or solifluction sheets. However, some concept of the mode of origin of rock glaciers is necessary to an appreciation of the climatic factors involved in the formation of the rubbles. The first description of forms similar to rock glaciers was made by J. G. Andersson in 1906 (pp. 91–112). He ascribed them to solifluction and assumed that they had originally been composed largely of fine debris, which was later removed. In 1910 Howe published a paper on the landslides of the San Juan Mountains. He described many rock glaciers, under the name "rock streams," and ascribed to them a landslide origin. Evidence presented by Capps (1910, pp. 371–375), Patton (1910), Tyrrell (1910, pp. 550–551), Chaix (1923, pp. 10–19), and Kesseli (1941, pp. 210–215) shows clearly that they are not the work of landslides. Capps and Moffit (1911, p. 56) believe that rock-glacier movement proceeds in a glacier-like way and is due to movement of interstitial ice, although no doubt all true glacial ice has disappeared. Chaix and Kesseli expressed the opinion that they originate in the waning stages of glaciation as the ground moraine of an overloaded and enfeebled glacier and that they continue their motion by some process of frost heaving or solifluction after true glaciers have disappeared from the region. Chaix measured the rate of movement of some rock glaciers in Switzerland and found it to amount to more than 1 meter per year. Tyrrell, on the other hand, holds, with Patton and Howe, that their origin is not solely through glaciation. He cites as example a rock glacier on the outskirts of the city of Dawson, Yukon Territory, outside the region of continental glaciation. This rock glacier rises at the base of a cliff of greenstone, from which a spring emerges. In the winter this spring fills the rock glacier with a mass of ice, and it is to movement induced or aided by this ice that Tyrrell ascribes the advance of the mass of greenstone blocks over gentle terrain. The general appearance of the younger of the rubble deposits, as described below, indicates that they have moved in a manner closely similar to that of rock glaciers; yet their existence outside the limits of glaciation and on a mountain below the lower limit of the accumulation of Wisconsin time indicates that they have an origin independent of true glaciers.

MODERN TALUS, FROST FORMS, AND THE YOUNGEST RUBBLE DEPOSITS

(FIFTH STAGE)

With the exception of a few outcrops, the steep slopes of Jumbo Dome are entirely mantled with sheets of detritus (fig. 3). This detritus consists of angular blocks of andesite, equidimensional or slablike in form, averaging 2–4 feet in diameter, and ranging up to 15 feet in greatest diameter. These sheets of coarse detritus are derived from the underlying rock by strong frost action and move down slopes partly by rolling and slipping and partly by creep due to freezing of interstitial water. They resemble the incipient forms of rock glaciers.

Most of the rock fragments or blocks are covered with a mantle of dark lichen on all exposed surfaces, whereas bottom surfaces or those in contact with other boulders are bare. The general aspect of the slopes is therefore very dark, almost black. According to H. M. Raup[2] the development of such a lichen cover requires at least two hundred years, the minimum

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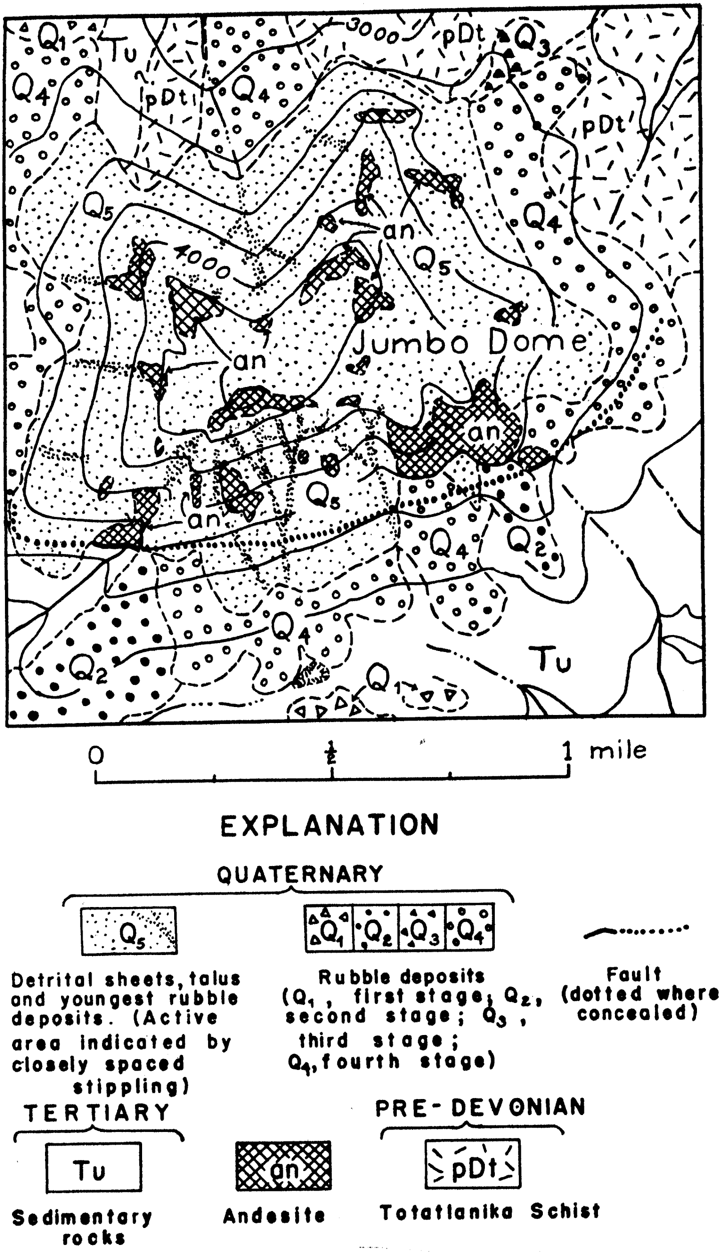
[2] Personal communication.


Q1
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Q5
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0
½
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EXPLANATION
QUATERNARY
Q5
Detrital sheets, talus and youngest rubble deposits. (Active area indicated by closely spaced stippling)
Q1 Q2 Q3 Q4
Rubble deposits (Q1, first stage; Q2, second stage; Q3, third stage; Q4, fourth stage)
Fault (dotted where concealed)
TERTIARY
Tu
Sedimentary rocks
an
Andesite
PRE-DEVONIAN
pDt
Totatlanika Schist

FIG. 3.—Geologic map of Jumbo Dome, Alaska

period for which the sheets of lichen-covered rock fragments have been stable.

The fragments of a few areas are blocks of fresh andesite, having a light pink color. Presumably, the movement of blocks in these places is active (pl. 1, *B*). The limited areas of active movement are located (1) below areas of exposed bedrock from which newly frost-riven blocks fall or move down slopes; (2) in places where the average grain size of the detrital sheet is less than 4 inches and some fine material is present; (3) on steep detrital cones where meltwater runs off, forming torrent levees. These detrital cones usually occur below re-entrants and amphitheaters in the higher slopes. The torrent levees, which are double parallel ridges of boulders, lead downstream from the re-entrants. At the lower ends of the torrent levees, the groove between them commonly passes into a steep-walled stream channel, the bottom of which is 10 or 12 feet below the surrounding slope. The torrent levees and the channels which extend down the slope from their lower ends are due to the action of meltwater in the spring and probably of summer cloudbursts. Water can be heard running beneath the grooves between the torrent levees, and the channels downslope from them have a small flow of water at the surface during most of the summer. Forms of somewhat similar appearance, described by Sharp (1942, pp. 222–227) from the Saint Elias Range, end in masses of bouldery mud and are apparently levees of alpine mudflows. Matthes (1930, pp. 108–109) has described similar forms from Yosemite Valley and attributes them to torrent action. The form and nature of the active talus deposits, involving as they do finer debris and running water, are not suggestive of an arctic or glacial climate as are the forms described in the following paragraphs.

In the upper part of the south face of the mountain a series of narrow horizontal curving bands of fresh blocks suggests a slight amount of movement en masse of the detrital sheet (pl. 1, *B*), which is probably an incipient rock glacier. The undisturbed and lichen-covered nature of the greater part of the block-covered slope indicates that the movement has affected only a small portion of the slope.

Differential movement of parts of the detrital sheet on the east side of the mountain is shown by lobelike masses of blocks, about 60 feet long down the slope, 20–40 feet wide, and rising 3–5 feet above the slope. The platy or flatter rock fragments are characteristically imbricated and dip upslope in the central parts of the lobes. Along the sides of the lobes they are oriented parallel to the direction of slope and dip inward beneath the lobe. Along the front of the lobe they are oriented parallel to the contour lines and dip beneath the lobe. The boulders of these lobes are lichen-covered and, therefore, have probably been more or less stable for at least two hundred years. However, rock glaciers have been observed where the boulders are lichen-covered but where overturned trees, still living, show that some movement still continues.[3]

The detrital sheet on the north slope of the northeast corner of Jumbo Dome has a series of low transverse ridges of detritus which have the appearance, seen from a distance, of gigantic ripple marks. The ridges are about 100 feet apart, from crest to crest, and appear to pitch 30°–45° to the west, down the slope (fig. 4). The ridges were not examined close at hand,

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[3] S. R. Capps, personal communication.

and their height above the slope is not known. They are made of the same coarse debris that makes up the detrital sheet. Each of them terminates upward both sharply and at a uniform level. The process of their formation is not known, but it is believed that intense frost action, of the type that produces the lobelike masses on the east slope, is the major agent. The detrital sheet on which the ridges are found is dark colored and is probably covered with lichen, although other plants are probably scarce or absent. The ripple-like ridges probably were formed at the same time as the lobelike masses on the east side of the mountain.

From the base of the steep detrital sheet on the northwest slope of the mountain, well-developed masses of rubble which resemble rock glaciers have advanced over older rubble deposits on a slope of less than 15°. These rubble deposits extend down to an altitude of 3,300 feet and are thickest at their lower ends. They are differentiated from older rubbles by the absence of fine detritus at the surface and by the absence of any plant life other than lichens on them. They rise above older rubbles with a steep slope. A side view of a typical mass of the youngest rubble (fifth stage) advancing over older rubbles is shown in plate 2, *A*. The size and shape of this deposit of rubble resembles in many respects an incipient rock glacier, although it does not occur in a cirque. The presence of lichen on the boulders of these youngest rubbles suggests that they have been more or less stable for two hundred years and were formed in a very recent epoch of more rigorous climate.

RUBBLE DEPOSITS OF THE FOURTH STAGE

The youngest rubbles (fifth stage) rest on older rubble derived from Jumbo Dome. Like the younger rubbles, the older deposits consist largely of angular blocks of andesite, but a considerable admixture of fine materials derived from the weathering of andesite blocks is also present. This older rubble was carried downslope in a manner resembling the movement of rock glaciers and is attributed to the fourth recognizable stage of rubble movement on the mountain.

Rubble of the fourth stage can be distinguished from the youngest rubbles and detrital sheets by the degree of weathering it has undergone and by its cover of vegetation, which the rubble deposits of the fifth stage lack. That this difference in vegetative cover is the result of recency of movement of the youngest rubbles and not of altitudinal changes is demonstrated by the fact that the flat upper surface of Jumbo Dome bears a luxuriant tundra-type vegetation. On the surface of rubble deposits of the fourth stage are numerous patches of poorly to well-developed polygonal soils: stone rings, isolated silt patches, and stone stripes.[4] Where these are present, there is no vegetation. These modern frost phenomena are thought to develop on top of the older rubble deposits and to represent the extent of present-day frost modification. Many of the andesite blocks have been split by frost action

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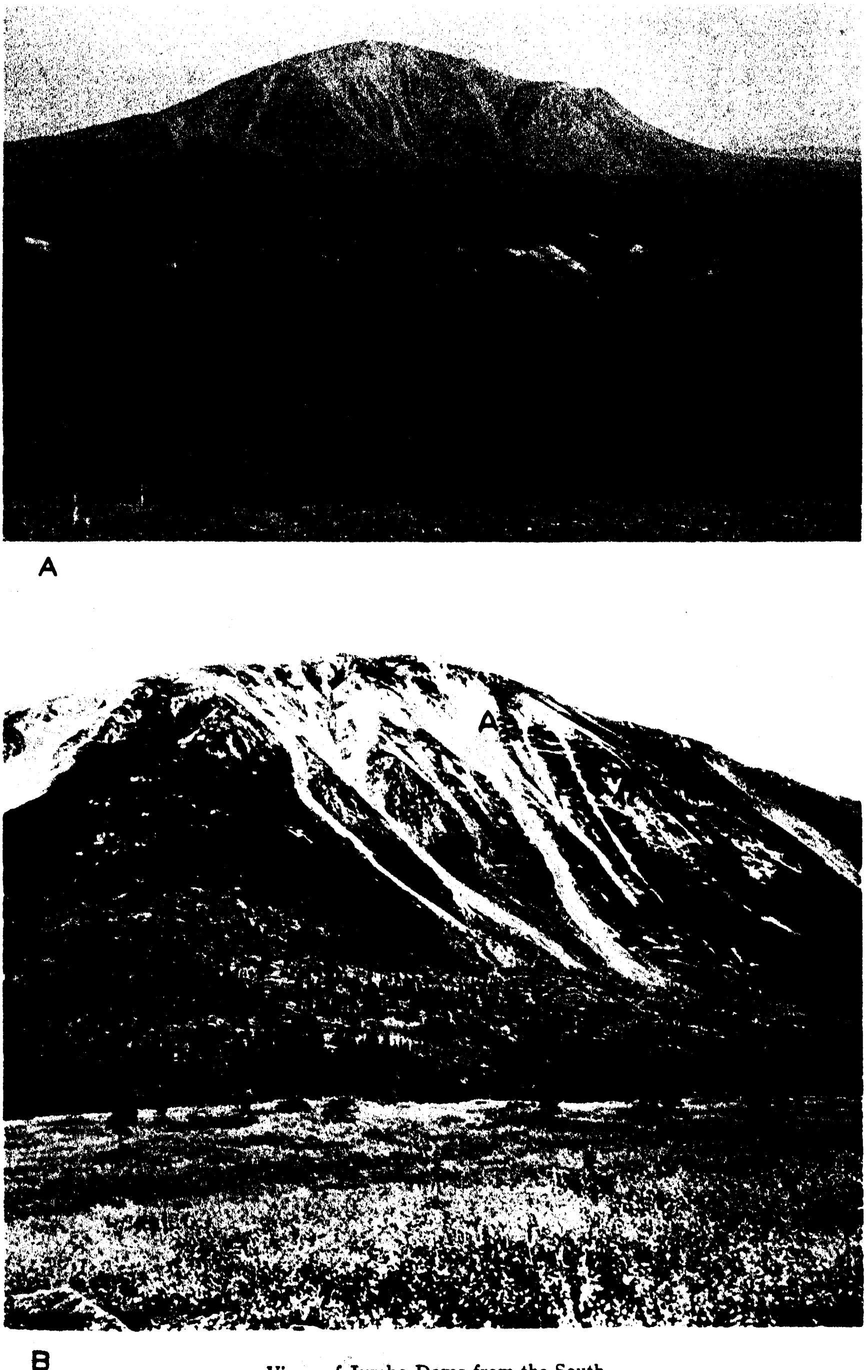
[4] For definition of these terms see Sharp (1942*b*).

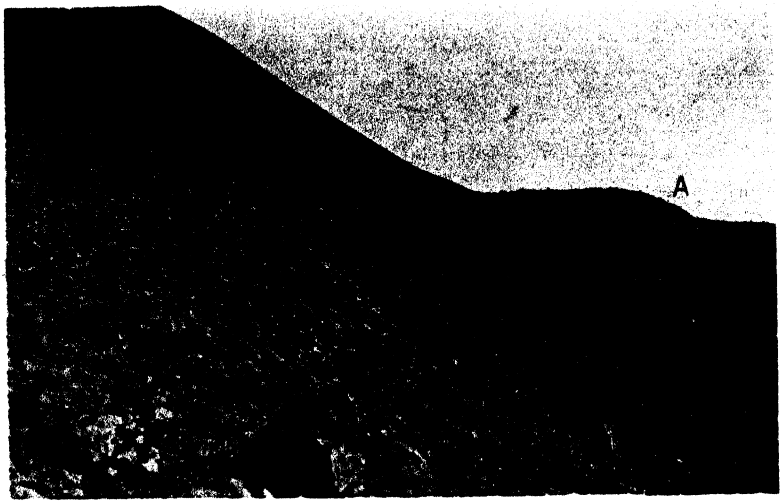
PLATE 1

A, View of Jumbo Dome from the south.

B, South face of Jumbo Dome, from hill of first-period rubble to the south. Active areas on the detritus-covered slope are light; stabilized, lichen-covered areas are dark. Note horizontal bands of incipient rock-glacier movement near *A*.



Views of Jumbo Dome from the South



A

B

Rubble deposits and badlands

into smaller fragments, which remain in the positions that they had in the original block, although spread by cracks 1–3 inches wide. All exposed surfaces of these fragments, including the walls of the cracks, are covered by a thick coat of black lichens, as are all other blocks on the rubbles of the fourth stage. This frost-splitting is the modification of the fourth-stage rubbles which formed during the development of the fifth-stage rubbles.

In general, rubble deposits of the fourth stage have well-preserved looped ridges, which are brought out by unequal growth of vegetation on the ridges and in the intervening depressions. They (1) have distinct frontal slopes, which are usually modified somewhat by creep; (2) display convex upper surfaces; and (3) resemble rock glaciers. In general, they appear relatively little affected by subsequent stream erosion, although in a few places they are channeled by gullies 15 feet deep. The largest rubble deposits of the fourth stage are found on the north and west sides of the mountain, where they extend down valleys to an altitude

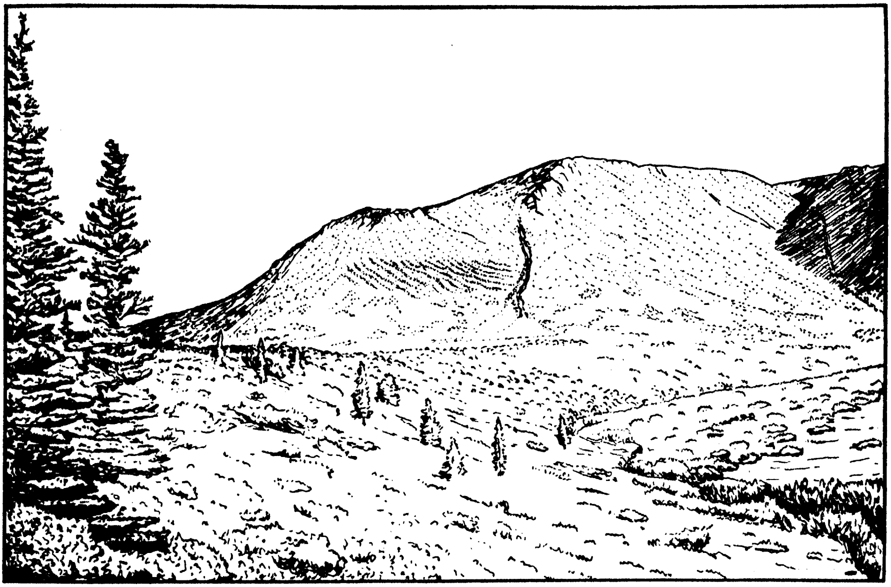


FIG. 4.—View of Jumbo Dome from the north (sketched from a photograph). Ripple-like frost forms can be seen immediately below the left-hand point of rock in the background; to the right of these is a small gully lined with torrent levees.

PLATE 2

A, Toe of rubble deposit, probably a rock glacier, of the youngest (fifth) stage (*A* in photograph), resting on older frost-moved rubble, at northwest corner of Jumbo Dome, from the north. Rubble of fourth stage in foreground.

B, Badland area in basal Tertiary conglomerate at head of McAdam Creek, 4 miles northeast of Jumbo Dome. The gentle surface in the middle background (*A* in photograph) is underlain by 20 feet of solifluction debris derived from the mountains of Totatlanika schist (*B*) in the far background.

of 2,500 feet, to about 700–800 feet below the lower limit of rubble deposits of the youngest (fifth) stage.

RUBBLE DEPOSITS OF THE THIRD STAGE

Angular andesite boulders cover the floors of valleys draining Jumbo Dome on the north and west from an altitude of about 2,250 feet upstream to the lower ends of the rubble deposits of the fourth stage. These masses of angular debris have no topographic expression, such as is shown by the younger rubble deposits. They do not rise above their surroundings but merge with them. However, as these sheets and masses of coarse debris consist of andesite blocks which have been carried 2,500–6,000 feet horizontally and 500–1,200 feet vertically below the contact between the andesite and sandstone, it is presumed that they were once rock glaciers or some related type of deposit which owed its movement to processes active under a climate colder than the present one. The rubble deposits of the third stage on the north side of Jumbo Dome are cut by gullies which are choked by rubbles of the fourth stage. The absence of any topographic expression of the rubbles of the third stage, such as is characteristic of the rubbles of the fourth and fifth stages, suggests that the former are probably of considerably greater antiquity than the latter.

RUBBLE DEPOSITS OF THE SECOND STAGE

Scattered blocks of andesite forming a thin sheet of rubble cap a ridge of Tertiary rocks that extends north from the prominent point on the northeast corner of Jumbo Dome (fig. 2). This ridge rises about 50 feet above the surrounding valleys, which are filled with rubbles of the third and fourth stages. Between these scattered boulders and the mountain from which they were derived is a stretch of ridge about 2,000 feet long, on which Tertiary sandstone and Paleozoic(?) schist crop out. The boulders have thus been moved 2,000–3,000 feet horizontally on a slope of less than 15° and presumably are the remains of a formerly more extensive rubble deposit. The scattered boulders were brought to their present position from the mountain on a surface into which stream valleys have been incised at least 50 feet, and probably more. This thin dissected sheet of detritus represents a period of rubble formation much older than the third, fourth, or fifth stage and separated from them by a considerable period of erosion.

Rubbles of andesitic debris, capping the outcrops of Tertiary sandstone exposed in steep-sided badland-like gullies on the southwestern and southeastern corners of Jumbo Dome, are tentatively correlated with the second stage. They are separated in part from their original source (the shoulders of the mountain) by gullies comparable in size to the gullies which dissect the rubble deposits of the second stage on the north side of Jumbo Dome. On the other hand, these second-stage rubbles descended into valleys which had been carved in rubbles remaining from a still earlier stage (first stage). In addition, the second-stage rubbles appear to have been derived, in part, from rubbles of the first stage.

RUBBLE DEPOSITS OF THE FIRST STAGE

Deposits of angular boulders of andesite similar to those which are described above are perched on the ends of spurs extending from Jumbo Dome and on the tops and sides of hills now separated from Jumbo Dome by broad valleys 100 feet deep. Some of these deposits are as much as 1½ miles from Jumbo Dome and must have been transported on a slope of less than 15°. They are remnants of an apron of rubble extending from Jumbo Dome. The apron originally must have been so extensive that presumably a vast field of blocks extended from the dome on all sides as far as 1–2 miles from the contact between the andesite and sandstone. These rubbles are older than all the others and are assigned to the first stage of rubble formation. Rubbles of the younger stages occupy the valleys which separate these rubble deposits of the first stage from Jumbo Dome. Where first-stage rubbles are found on the ends of ridges extending from Jumbo Dome, younger rubbles, having ridden out on these ridges, rest on exposed Tertiary rocks which separate the oldest rubble deposits from their source. This is clearly seen on the ridge that forms the west side of the canyon of Winter Creek (fig. 2). Here rubble deposits of the third and fourth stages rest on Tertiary rocks which form the crest of the ridge. The triangular north slope of this ridge is covered with large angular blocks of andesite which are almost completely buried in the mossy tundra vegetation which covers this ridge. Exposures on the east and west walls of this ridge indicate that these blocks form a veneer over north-dipping Tertiary rocks, which, in turn, form a plate about 100–150 feet thick, resting on schist.

Similar rubble deposits of coarse angular andesite blocks form the noses of the northwest, west, and southwest shoulders of the mountain. On the ridges behind each of these deposits Tertiary sandstone comes to the surface, and the rubbles are completely dismembered from their source. The canyons between these ridges, which contain rubble deposits of the third and fourth stages, appear to have been cut after the older rubble deposits were formed.

South of Jumbo Dome and north of Marguerite Creek is a haphazard jumble of irregular hills, in part flat-topped, in part steep-sided and with narrow crests, which are separated largely by undrained depressions, a few of which contain ponds. At first glance this appears to be landslide topography, and the possibility that a large landslide was once precipitated from the south face of Jumbo Dome was considered. However, most of the steep-sided hills consist of Tertiary coal-bearing rocks in place, and the undrained depressions lie in intervening valleys which are choked by rubble of angular andesite blocks. The broader flat-topped hills are covered by a sheet of rubble consisting of andesite blocks and differing in no way from other rubble sheets. From these flat-topped hills streams of blocks extend down the slopes eroded in Tertiary clays and sandstones. These block streams are now completely covered with moss and are therefore ancient. They were not active as late as the time of formation of the rubble deposits of the fifth stage. The explanation invoked to account for this curious assemblage of hills is that a field of rubble once extended from Jumbo Dome as a broad apron covering a gentle surface, which has a present altitude of about 3,000 feet (fig. 5, *A*). This apron presumably was formed under the influence of the cold climate of a glacial stage or substage. A period of stream erosion, probably related to a warmer interglacial or interstadial climate, followed the development of this rubble; canyons a few hundred feet deep were cut into it from the south, east, and west. Canyon walls in the soft Tertiary rocks had steep, bare slopes, such as exist in recent badlands formed by tributaries to Lignite Creek, a few miles south. As the streams dissected this rubble apron, isolated steep-sided hills capped with coarse andesite blocks

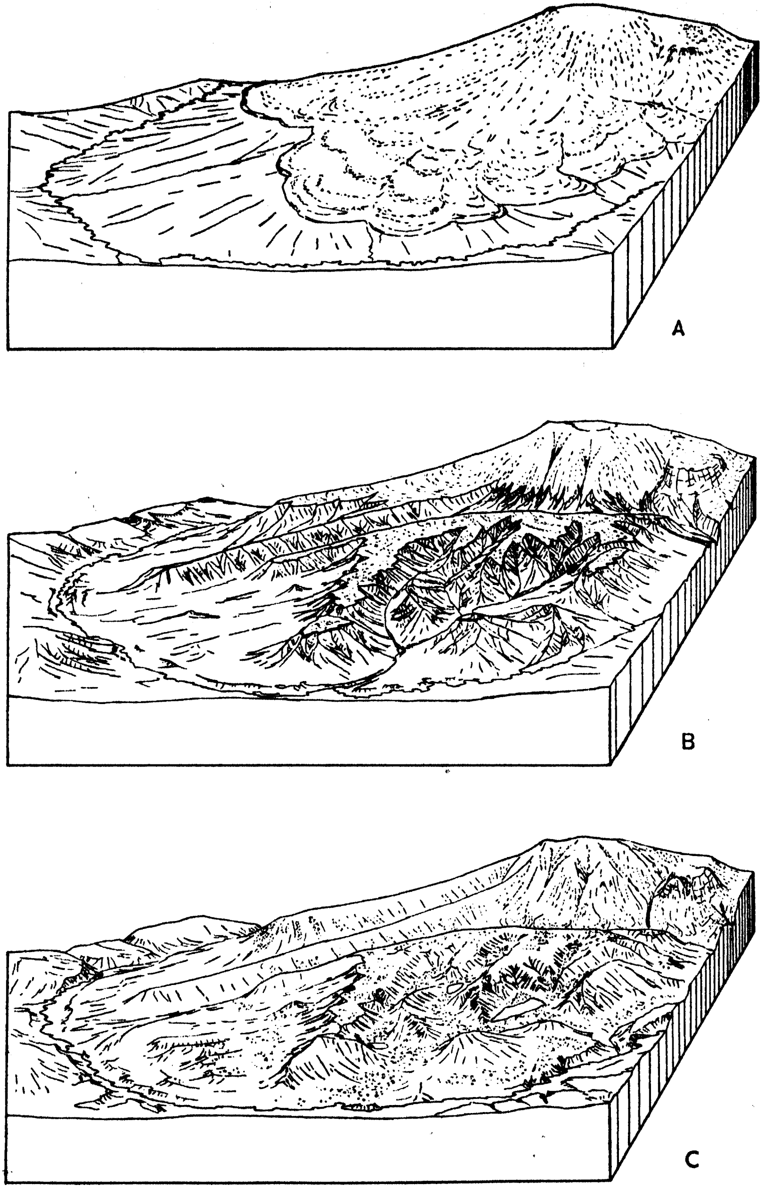


FIG. 5.—Block diagrams representing the development of topography on the south side of Jumbo Dome. *A*, south side of Jumbo Dome at close of first stage of rubble formation. *B*, south side of Jumbo Dome at close of period of badland-forming erosion following the first stage of rubble formation. *C*, present appearance of the south side of Jumbo Dome, after later periods of frost action and rock-glacier movement.

must have dominated the local topography. The appearance of the area at that time is diagrammatically illustrated in figure 5, *B*. Another change in climate brought a return to conditions favoring mass movement of blocky material and unfavorable to canyon-cutting. The masses of angular blocks slid off the hills and into the narrow canyons, filling them to a greater depth in some places than in others and producing a peculiarly haphazard-looking topography, with numerous undrained depressions, which is shown diagrammatically in figure 5, *C*.

It is likely that more than one stage of canyon-cutting and hillside-sloughing is involved in the destruction of the original broad apron of rubble. In the short time spent in the area, however, it was not possible to establish criteria for more than one period of stream erosion succeeded by one period of intensive frost action. The history, as represented here and illustrated in figure 5, is to be regarded as a simplification of a presumably more complex story. This group of hills is separated from Jumbo Dome by a valley into which deposits of rubble, tentatively correlated with the second stage, have spilled. It is therefore older than the second stage and is tentatively correlated with the first stage. Because this broad apron of rubble is closer in altitude to the present stream grade than are other deposits assigned to the first stage, it may represent a separate stage. It was not possible, however, to develop a satisfactory criterion for distinguishing it from other deposits of the first stage; it is therefore grouped with them.

INFLUENCE OF CLIMATE ON DENUDATION IN THE ALASKA RANGE

It has been shown above that the rubble deposits and the detritus-covered surface of Jumbo Dome are, for the most part, inactive and that there probably has been no significant movement of the most recent rubble deposits for at least two hundred years. Such movement as has been observed is related to less rigorous climatic conditions than those which prevailed during the formation of the rubble deposits. Soil features and surface forms of the type described by Högbom (1914) and Sharp (1942*a*), among others, and ascribed by nearly all workers to factors introduced or intensified by cold climates show a strong altitudinal zoning of the lower limits of their present activity on the north side of the Alaska Range. This zoning is related largely to the size and degree of sorting of constituent particles of the soil involved in the movement. In the vicinity of Jumbo Dome, the lower limit of presently active mass movement of debris, consisting predominately of blocks larger than 1 foot in average diameter, is about 4,600–4,800 feet (or just above the summit of Jumbo Dome). Polygonal soils in poorly sorted, moderately fine debris are active above an altitude of 2,300 feet, and solifluction or creep is taking place above an altitude of 2,000 feet, beneath moss and tundra cover on some north-facing slopes in exceptionally favorable situations and where underlain by Tertiary sandstone and claystone.

The rapid downslope movement of debris under conditions of rigorous climate results in overloading the streams and causes them to aggrade their beds. This has been recognized by students of the Pleistocene in Germany (Zeuner, 1945, pp. 25–27) and in the Middle West (Lueninghoener, 1947). Matthes (1900, pp. 182–184) reported this effect in the Big Horn Mountains. This phenomenon also is clearly demonstrated in many of the higher, unglaciated parts of the Alaska Range, such as the headwaters of the Savage River and of Healy Creek. The constant downslope mass movement tends to fill in gullies as rapidly as they are formed and inhibits many characteristic types of stream erosion, such as badland formation, gullying, etc. The results of such a degradational regime are smooth, featureless hillslopes of large extent, such as are found throughout much of interior Alaska. Only through a climatic amelioration sufficient to slow considerably or completely to stop mass movement of soils can normal stream erosion become dominant. Stream erosion is generally dominant today below an altitude of 4,000 feet. This is amply attested by terrace-cutting, gullying, and badland erosion that are active everywhere below this altitude throughout the north flank of the Alaska Range. Badlands are restricted almost entirely to the Tertiary formations. Bare, actively eroded badlands occupy only a small portion of the surface area on which one would expect to find them in this region. Typical of these is the badland shown in plate 2, *B;* this area is in typical basal Tertiary conglomerate and is located about 4 miles northeast of Jumbo Dome, at an altitude of 2,500 feet. Here active dissection is destroying an older gentle surface (seen in right background) mantled with a layer of solifluction debris 10 feet thick. Badland erosion is initiated through landslides which take place during the thawing-out of the upper layer of frozen ground. This lays bare the underlying soft sandstone and shale, which are readily attacked by meltwater in spring and by the runoff of summer thunder showers and autumn rainstorms.

Areas of small hills, of steepness equal to that of the badlands but covered with tundra-type vegetation and possessing rounded tops, are thought to be badlands formed in an earlier cycle of stream erosion and later stabilized through development of a turf cover during a cold part of the climatic cycle.

It is likely that the upper limit of terracing and badland formation always lies below the lower limit of movement of coarse rock debris, and it probably is as much as 500 or 1,000 feet below the lower limit of activity of bouldery rubbles or rock glaciers. It is evident that to bring about conditions for the erosion of rubbles by streams requires raising the isothermal surfaces at least to their present position. It was shown in an earlier part of this paper that the lowest altitude of peaks that have glacier-filled cirques is 7,200 feet and that of peaks bearing cirques of the Wisconsin stage is 6,000 feet. The lowering of isothermal surfaces necessary to produce rock glaciers on Jumbo Dome is, therefore, approximately that required to bring about conditions of the maximum of Wisconsin glaciation. Each period of rock-glacier formation may well represent a period of advance of the glaciers in the mountains corresponding to continental glaciation in other parts of North America. Correspondingly, each period of terrace-cutting, badland erosion, and cessation of activity of the rock glaciers should correspond to an interglacial or interstadial fluctuation of climate in North America.

THE PLEISTOCENE CHRONOLOGY INDICATED BY THE RUBBLE DEPOSITS OF JUMBO DOME

The evidence of the rubble deposits of Jumbo Dome suggests that the older glaciation which has been recognized in the Alaska Range was followed by at least five episodes of glacial climate, during which the movement of rubble on Jumbo Dome took place. The episodes were separated by intervals of less severe climate, during which the rubble deposits were dissected or otherwise modified. The two earlier of these episodes of glacial climate (represented by rubble deposits of the first and second periods of rubble movement) were separated from each other, as well as from both the older glaciation and the third period of rubble movement, by relatively long periods of climatic amelioration, during which considerable uplift and dissection took place. The three younger periods of rubble movement were separated from one another and from the present by much shorter periods of climatic amelioration, in which very little dissection and probably little uplift took place.

No attempt is made in this paper to correlate the episodes of rubble movement on Jumbo Dome with the glacial stages and substages of eastern North America. The short amount of time spent on the problem and the uncertainties involved in such a long-range correlation make any such attempt of doubtful validity. More field work in Alaska and in western North America undoubtedly will show that at least some of the periods of glacial climate indicated by the rubble deposits of Jumbo Dome correlate with glacial stages and substages in the United States.

Acknowledgments.—The manuscript was critically read and edited by Kirk Bryan, of Harvard University, and by Fred H. Moffit, S. R. Capps, and Charles S. Denny, of the United States Geological Survey, who have offered many helpful comments and suggestions. The writer especially wishes to acknowledge his indebtedness to Professor Bryan, without whose instruction these features would have been overlooked in the field. The writer also wishes to acknowledge the kindness of J. D. Henney and Nels Jackson of Ferry, Alaska, in permitting the use of their cabins. George C. Turner and Harrel C. Stacy were field assistants during the field season of 1947, when the area was mapped. W. S. Benninghoff, botanist, of the Geological Survey visited the area with the writer on September 17–20, 1948.

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REVIEWS

"Strukturböden, Solifluktion und Frostklimate der Erde." By **Carl Troll**. (*Geologische Rundschau,* Band 34, Heft 7/8.) Pp. 545–694. 1944.

The issue in which this paper occurs is named "Diluvial-Geologie und Klima" and is composed of an introduction and 13 papers, totaling 470 pages. The papers, which were assembled and edited by Professor Carl Troll, form an extraordinary contribution to the knowledge of the Ice Age. The list of the papers which follows shows the wide range of the subjects that are discussed.

Introduction. **Troll, Carl.** "Diluvialgeologie und Klima," pp. 307–325.
1. **Machatschek, Fritz.** "Diluviale Hebung und eiszeitliche Schneegrenzdepression," pp. 327–341.
2. **Pfannenstiel, Max.** "Die diluvialen Entwicklungsstadien und die Urgeschichte von Dardanellen, Marmarameer und Bosporus," pp. 342–434.
3. **Jaranoff, Dimitri.** "Das Klima des Mittelmeergebietes während des Pliocäns und des Quartärs," pp. 435–446.
4. **Louis, Herbert.** "Die Spuren eiszeitlicher Vergletscherung in Anatolien," pp. 447–481.
5. **Büdel, Julius.** "Die morphologischen Wirkungen des Eiszeitklimas im gletscherfreien Gebiet," pp. 482–519.
6. **Steeger, Albert.** "Diluviale Bodenfrosterscheinungen am Niederrhein," pp. 520–538.
7. **Weinberger, L.** "Frostspalten und Froststrukturen in Schottern bei Leipzig," pp. 539–544.
8. **Troll, Carl.** "Strukturböden, Solifluktion und Frostklimate der Erde," pp. 545–694.
9. **Aario, Leo.** "Die spätglaziale Entwicklung der Vegetation und des Klimas in Finnland," pp. 695–704.
10. **Finsterwalder, Richard.** "Eishaushalt von Gletschern und Niederschläge in Gletschergebieten," pp. 705–712.
11. **Wundt, Walter.** "Die Mitwirkung der Erdbahnelemente bei der Entstehung der Eiszeiten," pp. 713–747.
12. **Meinardus, Wilhelm.** "Zum Kanon der Erdbestrahlung," pp. 748–762.
13. **Behrmann, Walter.** "Das Klima der Präglazialzeit auf der Erde," pp. 763–776.

Papers Nos. 8 and 5 will be reviewed in this issue; others will be reviewed later.

The present paper by Professor Troll aims to distinguish climatic types of ground frost and of frost-caused earth forms (p. 672). Since the frost-induced or frost-facilitated denudation (erosion by extensive surficial mass movements), especially the solifluction, proceeds differently in different frost climates, knowledge of the modern processes enables climatic conclusions from past frost-caused earth forms.

Troll, who has studied frost action in the Alps, Scandinavia, the Andes, and the mountains of East and South Africa, starts from the contrast in climate and in frost-caused earth forms that exists between the polar regions and the alpine zone of tropical mountains. In the former the annual temperature variations are very prominent, the daily fluctuations insignificant, and the climate seasonal. In the latter the daily temperature changes are important, and the annual are almost or entirely absent, producing a day-and-night climate. In the polar regions there are large earth patterns; in the tropical mountains beautiful miniature ones.

Among kinds of ground frost there are: perpetual (permafrost, pergelisol), winter, short-periodic, and nightly ground frost, and also needle ice (brush ice) and lens ice (spring ice). The relationship of ground frost to the ground that thaws and to the ground that never freezes and the withdrawal of water from the substratum by surface freezing are important conditions and phenomena.

A genetic classification of frost-caused earth forms requires consideration of the physical composition of the material, the ground slope, and the water supply (duration of snow cover, depth and time of thaw) (p. 620). The role of the ground frost varies for different kinds of earth forms. The increase in volume with freezing produces pressure. Melting of the ice causes supersaturation of the ground and facilitates its displacement (p. 550). Thus a satisfactory genetic classification is not yet possible, but Troll attempts one (pp. 547–550, 620, 675). Before presenting this, it is necessary to find appropriate American names and terms. "Earth" as in "earth dam" is suggested, not "soil," which is composed of mineral and organic materials and which is lacking or poorly developed in the principal regions of frost action. Stone rings, stone stripes, etc., are not different kinds of earth but different patterns of, and in, frost-separated earth. Such expressions as "earth structure," "structure earth," "earth texture," and "texture earth" are dubious and avoided, because they may be confused with the well-established terms "soil structure" and "soil texture." "Soil structure" has reference to the morphological aggregates in which the individual soil particles are arranged (prismatic, columnar, massive, etc., structure). "Soil texture" refers to the relative proportion of the various size groups of individual soil grains (sandy, silty, clayey, etc., texture).

Troll's classification of the principal frost-caused earth forms (*Frostbodenformen*) is as follows:

A. Frost-caused earth patterns (*Frostmusterböden*, *Frostgefügeböden*, or *differenzierte Frostbodenformen*)
  1. Frost-caused earth patterns in heterogeneous material (*Strukturböden*)
    a) On level ground: stone rings, stone nets, earth islands
    b) On slopes: stone stripes, earth stripes
  2. Frost-caused earth patterns in homogeneous material (*Texturböden*): fissure nets or fissure polygons, ice-wedge fissure nets, or tundra polygons

B. Frost-caused amorphous earth forms (*amorphe Frostböden*)
  1. Vegetation, soil in true sense, and/or peat, important contributing factors
    a) On level ground: sod mounds or turf hummocks, earth hummocks, miniature peat mounds, peat mounds or *Palsen*, block fields, ice-lens knolls or spring-ice knolls (*Aufeishügel*)
    b) On slopes: turf-banked terraces or benches, solifluction ridges and tongues, stone garlands, stone or block streams, string moors
  2. Snow pressure, snow sliding, wind, etc., contributing factors: earth plaster or mortar, sod lumps (*Rasenschälen*)

True frost-caused earth patterns occur at certain elevations in all climatic belts of the globe but are best developed in the polar regions and in high tropical mountains, and poorest in the mountains of the mid-latitudes (pp. 551–557, 673). Their climatic lower limit rises and falls with timber line and snow line, and it rises toward the Equator and the interior of mountain ranges and also with increasing continentality and generally lies above the modern timber line and the glacial-age snow line.

Frost-caused earth patterns are thus subnivale earth forms and require only a periodic, strong, and morphologically active ground frost (pp. 557–562, 674). A long and severe winter frost can have similar effects as pergelisol, i.e., can give rise to large earth patterns. Nightly frost and daily thaw throughout the year in the alpine zone of tropical mountains produce in water-bearing earths an excellent assortment of the debris into miniature patterns which completely lack vegetation. In latitudes between the polar type and the tropical type of ground frost there are transitional types and mixtures.

Several fundamental problems concerning the formation of frost-caused earth forms are discussed (pp. 562–575, 674). The controversy between the convection and involution (Brodel) hypotheses, on the one hand, and the frost-pressure hypothesis, on the other, is not to be decided exclusively in favor of either. Frost pressure, consisting of vertical frost heave and horizontal frost shove, is of prime importance—but in high latitudes at least the earth can become so supersaturated with water from excessive ground-ice that it assumes the character of a suspension. Frequency of temperature fluctuations about the freezing-point is of prime importance to the miniature earth patterns of low latitudes, while it is of subordinate importance in high latitudes with cold winters.

The concept of solifluction is broadened to include not only earth flow over a perpetually or seasonally frozen substratum but also sorting of the debris on level ground (e.g., formation of stone rings) and debris movements caused by short-periodic or daily alternation of freezing and thawing of the topmost earth layers. A type of day-and-night solifluction is needle-ice formation.

The forces producing the frost-caused earth forms include changes of temperature and of volume, cohesion of freezing ground water, gravity, and water absorption of colloidal materials, especially clay and humus.

Needle ice consists of dense bundles of needle-like ice crystals beneath the ground surface and at right angles to it (pp. 575–592, 675). The best descriptive name would be "brush ice" (*Bürsteneis*). The ice needles are usually covered by a veneer of raised earth and pebbles. Needle ice forms by brief, usually nightly, frosts in fine-textured ground that is more or less free from vegetation and snow. If the ice is not melted by day but can continue to grow for nights or weeks, the ice formations of the individual nights are marked by a distinct lamination, and the ice may ultimately attain a thickness of over a foot. When needle ice forms on sloping ground, the raised debris will, upon melting of the needles, drop or roll to positions farther down the slope. An important type of erosion caused by needle ice is sod peeling (*Rasenabschälung*), at which the ice excessively raises the edge of the sod, causing it to break off and fall apart. Therefore, needle ice carries out considerable morphological work; and it frequently collaborates with the wind, which removes raised particles. Needle ice is important in all latitudes except the polar regions and plays its greatest role in winter-dry subtropical mountains.

Under particularly favorable ground and moisture conditions distinct earth patterns are formed locally far outside the climatic limits of frost-caused earth patterns (pp. 592–600, 675). On Öland in the Baltic beautiful miniature stone nets are formed in shallow (⅛–1 foot), poorly-drained earth resting on a comparatively impervious limestone. The limestone fills the role of frozen ground, considerable upfreezing of stones and plants takes place, and the wet, soft earth is highly mobile during thaws.

In the alpine zone of tropical mountains there are formed strikingly regular miniature earth patterns caused by nightly freezing, viz., stone nets, stone stripes, and earth stripes (pp. 600–616, 675). A special type of stone nets with exceptional assortment, with sharply segregated stone-free clay fields, are the cake polygons (*polygonale Kurhenböden*). The clay cakes, which are 1–2 inches (3–5 cm.) thick, are frozen hard nightly and have become raised by frost heave. They are so firm and compact that they frequently protrude slightly. The never frozen soft and moist substratum is slightly lower under the cakes than beneath the stone meshes.

The frost-caused earth patterns in the polar regions, which are formed in summer-thawed earth above permanent frost, are mostly of large size (pp. 613–620, 676). Ice-wedge fissure polygons (tundra polygons) are sometimes gigantic. However, the large patterns sometimes contain small ones, and these, in turn, occasionally miniature forms. The large patterns may be caused by the seasonal course of freezing and thawing, the small and miniature ones by short-periodic and daily fluctuations about the freezing-point.

The greatest variation of frost-caused earth forms is to be found in the subarctic regions (pp. 620–648, 676). In the Scandinavian mountains there are large patterns of the polar type. In Iceland there are large patterns in the cold interior and miniature ones (tropical type) in the oceanic coastal regions, where the temperature frequently oscillates between frost and thaw. ***Thuvur***, sod-covered earth hummocks, are amorphous forms which are common in Iceland and also have a wide distribution in subarctic and cold temperate regions. In the subarctic tundras of Eurasia and America, especially in their most continental parts, ice-wedge fissure nets are important features. Typical of the tundra are also earth islands, earth hummocks, and peat mounds or *Palsen*. In the taiga there is the string moor, which is a peat bog with alternate ridges and troughs which follow the contours of the gently domed moor surface. Thermokarst or ground-ice karst, formed by differential melting of the ground-ice, is important in eastern Siberia, while ice-lens knolls or spring-ice knolls (*Aufeishügel*) are characteristic of both eastern Siberia and of winter-cold central Asia. The spring-ice knolls, which are 20–100 meters in diameter and 1–6 meters in height, are formed underground by freezing of rising spring water.

The frost-caused earth forms and their climatic significance in the mountains of the temperate zones are little known (pp. 648–661, 677). Those in the southern Urals and the Riesengebirge are essentially of the large type; those in the oceanic British Isles, of the miniature type. In New England there are modern miniature earth patterns and fossil large ones, both of which deserve a more thorough geologic-climatic study than has been accorded them. In the Alps are features of two classes. Below modern glaciers, where the ground is thoroughly wet, from 2,000 meters altitude upward, there are rather distinct earth patterns of the large, polar type. In the vicinity of névé fields there is frequently mosaic-like earth plaster or earth mortar. Otherwise, miniature patterns are characteristic for elevations above 2,700 meters. There are also earth hummocks, miniature peat mounds, stone garlands, stone-banked terraces, turf-banked terraces, etc.

Frost-induced earth patterns in subtropical mountains are largely of the miniature type (pp. 661–672, 678) and are found in western Asia (the Toros, Caucasus, Elbruz, Pamirs, etc.), Japan, South Africa, and Argentina. However, those in the monsoon mountains of northern India, of Tibet, China, and Manchuria, where there are heavy summer rains and severe winters, are of the large type. In the Drakensberg of South Africa the pattern formation is caused by nightly freezing during the largely snow-free winters and in the Andes of Argentina by nightly frost during the snow-free summers.

This comprehensive monograph brings the knowledge of frost processes and features a long step forward.

ERNST ANTEVS

"Die morphologischen Wirkungen des Eiszeitklimas im gletscherfreien Gebiet." By JULIUS BÜDEL. (*Geologische Rundschau*, Band 34, Heft 7/8.) Pp. 482–519. 1944.

In opposition to old views that the land forms of the unglaciated areas of central Europe have been largely shaped by stream action ever since the Tertiary, Büdel argues that the low temperatures of the glacial ages promoted other powerful geomorphic processes and that the surficial features thus produced still dominate the landscape, owing to small topographic changes during the Postglacial (pp. 483, 515). The main facts and observations on which these opinions and conclusions are based are as follows:

In the mountains of central Europe there is a thick mantle of debris, which, even on very gentle slopes, contains large boulders, miles from their sources. Intercalated loess and frost-caused earth patterns, as well as overlying peat, show that all movement occurred during, and ceased with, the tundra climate of the Last Glacial. The debris accumulated by means of solifluction (earth movement and sorting caused by alternate frost and thaw) (pp. 487, 490, 515).

The lower altitudinal limit of modern solifluction usually lies in the scrub-forest belt, a little above the tall timber (p. 491). It stands at 2,000 meters in the central Alps, 1,500 in the Riesengebirge, and as high as 1,000 meters in the continental Kebnekaise in Swedish Lapland. However, it drops quickly as the climate becomes oceanic and is found at the sea level in northernmost Norway and in Iceland, where the ground is wet and the temperature frequently fluctuates about the freezing-point.

During the glacial ages the lower limit of solifluction lay at sea level as far south as central Europe, and the solifluction deposits here are widespread and well preserved. They are also thick because of the deep summer thaw of the perpetually frozen ground. In comparison, the postglacial and modern denudation (surficially extensive erosion), which consists of needle-ice solifluction, wind erosion, and soil washing, is small and very shallow (p. 492).

Also, the geomorphic activity of streams changed and decreased from glacial to postglacial times (p. 494). The uppermost valleys in the mountains of central Europe have a veneer on their sides and in their center a thick fill which was once a moving solifluction stream. These draws were formed by corrasion and are of glacial age, as shown by overlying peat beds. Some of them have been gullied in postglacial time, others have not (p. 495).

Below the junction of two or more draws, the unsorted solifluction deposit grades into a stratified stream deposit (p. 495). The cross section of the valley changes so that the sides now form a distinct angle with the floor, which in this section is wide and slightly domed in the center because of strong lateral erosion associated with the large debris transportation. Again the glacial age of the valley is proved by peat deposits. The modern stream channel, which is a result of a modest entrenchment during the Postglacial, is located at the side of the valley and skips at regular intervals from one side to the other.

Such nonglacigene valleys of brooks and small rivers can be studied to the best advantage at the foot of the Alps, where they can be accurately dated (pp. 496, 507). Here they start as solifluction draws on the glaciofluvial High Terrace (of Riss age) and are independent of the last glaciers. They are one to several kilometers long, a few hundred to several hundred meters wide, and terminate on a level with the Low Terrace (of Würm age). In the Inn region they are steep-walled and 40 meters deep. The entire valleys must have been formed during the Last Glacial, the Würm (p. 497).

The large valley deposits required plenty of debris and great transporting power by the streams (p. 499). The material was supplied by cold-climate solifluction, and the transporting efficiency by concentrated runoff and floods during the snow-melting in spring, in spite of smaller precipitation than at present.

The valley floor, which is domed in cross section, consists of a string of extended gravel fans. These cones occupy the entire width of the valley, originate alternately near the right and the left bank, and have caused the stream to shift from one side to the other; and they were renewed downvalley during floods. The modern streams are mostly entrenched in small channels (p. 500).

On the High Terrace the upper valleys, which are formed by solifluction, are distinctly asymmetric, having a gentle (cultivated) west side and a steep (forested) east side (p. 501). The asymmetry is a consequence of differences in exposure which influenced the erosion during the Last Glacial (p. 502). The factors are believed to have been: (*a*) exceptionally great deposition of loess and snow in the lee of the west bank by west winds, (*b*) consequent excessive solifluction on the west side, and (*c*) running water. Asymmetry prevails only as far down the valley as solifluction was more effective than running water (p. 504).

The streams of the valleys originating on the High Terrace have accomplished practically no erosion during the Postglacial age (p. 507). Hence the activity of glacial and nonglacial streams changed in the same manner with the transition from glacial to postglacial climate. During the cold ages glacial streams were loaded with debris released by glaciers, nonglacial streams by solifluction debris; the material was transported by floods from melting ice and snow or from melting snow alone; and at the same time broad valleys were carved by lateral corrasion. During the Postglacial the debris supply was insignificant and the floods relatively small both within and outside once glaciated areas. As stated, the conclusions hold for brooks and small rivers (p. 507).

The contrast between cold-age and warm-age valley formation—deposition and lateral erosion during glacials, vertical erosion during interglacials and late glacials—must have prevailed also before the Last Glacial (pp. 509, 514, 517). In agreement herewith there are in several central European mountains and in the nonglaciated eastern Alps four nonglacigene terraces corresponding to the four main glacial ages. The upper end of the deposits lies at the point where the stream leaves the mountains. The enormous terraced deposits in the Vienna basin are of this kind, being of glacial age but nonglacigene, for they do not start at end moraines.

Nonglacial Pleistocene terraces, which reflect the climatic history as faithfully as do glacial ones, are very important features in central Europe (p. 517).

ERNST ANTEVS

Études de cryopédologie. By ANDRÉ CAILLEUX. ("Expéditions polaires françaises, Missions Paul-Émile Victor.") Paris: Centre de Documentation Universitaire, 1948. Pp. 67; figs. 59.

Major features of frozen ground and former periglacial regions are described and figured in this outline summary of the literature. The subject matter is organized under the following subjects: soil temperatures, ground frost, ice wedges, polygonal and striped ground, soil plications and injections, block pavements, block streams, rock glaciers, debris patterns, turf mounds and corrugations, and mounds formed from ice lenses.

L. H.

"Landscape Study in South Africa." By LESTER C. KING. (Anniversary Address by the President.) (*Geol. Soc. South Africa Proc.*, vol. 50.) Johannesburg, 1947. Pp. xxiii–lii; pls. 2.

The work of King in Africa emphasizes the importance of back-wearing processes of surface denudation in contrast with peneplanation by down-wearing as manifested in other, more humid, regions. Like Bryan, for the American Southwest, and Johnson and Howard, for the Rocky Mountains, King rejects for much of Africa the Davisian process of peneplanation, in favor of pedimentation on a regional scale, or "pediplanation." It is by no means mere pedantry to insist on such a distinction; and it is not, as it might seem at first sight, a matter merely of interest to the student of geological processes. Not only do the commoner forms of relief in the African landscape cease to present a bizarre and exceptional picture in contrast with those of regions normally eroded, falling now into their places (equally "normal") in a systematic geomorphic scheme; but the denudation chronology of much of the continent bids fair to be immensely simplified, throughout a region in which the denudation chronology over a period exceeding one hundred million years must be relied on almost to the exclusion of the methods of stratigraphy for the elucidation of the geological history.

Some of the points noted by King as characteristic of African pediplanation are as follows:

> The three most important features of a landscape developed by pediplanation are:

> (1) The pediment or apron sloping outwards from a central hill mass towards the stream channels. This pediment *is continually being regraded* as the hill-core shrinks.
>
> (2) The process of slope retreat at constant, and usually steep, angle. The retention of this steepness is responsible for the abundance of scarps and steep-sided inselbergs in the African landscape.
>
> (3) The preservation, upon regions above scarps, of earlier erosion surfaces little modified by later weathering and erosion.

He notes also "several changes of outlook" which "follow adoption of the pediplanation concept":

> (1) As two surfaces, one above and one below the intervening scarp, are inherent in the theory, the old ideal of progressive approximation to a single planed surface loses much of its force.
>
> (2) Maturity is postponed indefinitely, owing to the long survival of initial surface, and indeed it only briefly precedes the ultimate stage of the pediplane.
>
> (3) . . . The retreat of slopes is at almost constant angle throughout the cycle.
>
> (4) Extensive alluvia are absent.
>
> (5) Whereas with peneplanation new cycles advance rapidly up the rivers and streams, and thus take on patterns either parallel to the main rivers or branching, with pedimentation as developed in Southern Africa, however, new cycles seem often to run but a short way up the rivers in advance of the main scarp between two cyclic surfaces. . . . The scarp is, therefore, much straighter in plan. . . .
>
> (6) It is also probable that as an agent of wide planation pedimentation is much more efficient and rapid than peneplanation in the original sense. At extreme pediplanation a landscape may be expected to reveal a multi-concavity upwards. Whether such a form, dominated by surface wash, continues to exist indefinitely, or whether the effects of weathering, apparent first as small convexities upon the crests of the interstream areas, gradually extend down the slopes to give a landscape of multiconvexities is yet uncertain. . . .

The concept of down-stepping benches, or multicycle peneplanation, as developed in Africa by Dixey, recognizes erosion cycles dated from late Jurassic, early Cretaceous, late Cretaceous, Miocene, Pliocene, and Pleistocene (two), and, as King notes, "each is deemed to have developed from the one preceding by erosion following uniform uplift." Dixey would explain the survival of very ancient surfaces as "due to the resistance to subsequent erosion of the hard formations across which they were cut" (to use his own words); but King, on the contrary, finds that "areas of greater altitude generally differ in no manner of such hardness from the lower areas encompassing them," and this he regards as consistent with the back-wearing and opposed to the down-wearing theory of denudation.

He remarks also that

> the postulate that a series of erosion steps such as distinguishes Southern Africa is necessarily due to diminishing intervals between uplifts of the subcontinent as a whole, is also due for reconsideration and the examination of possible alternatives. The pediplanation hypothesis alone does away with the necessity for such a postulate.
>
> The concept of five or six consecutive cycles being represented by restricted areas of planed surfaces at the heads of great river systems (regardless of divide migration . . .) is also due for revision.

This question is tied up with that of deformation of surfaces. "Dixey has claimed that many of the divides are surmounted by remnants of Cretaceous or Jurassic surfaces and are therefore ancient, even Mesozoic in age"; but King supports the views expressed earlier by Du Toit and Veatch that late warping of a "Miocene" surface has made the divides. King deduces with the aid of a diagram that the presence of residuals on divides is not inconsistent with a theory of their comparatively recent development by landscape warping. Well-planed areas on divides (such as the Congo-Zambezi divide) do not support a theory that the divides have been fixed in position for a very long time.

Warping is found to be more common than Dixey has been willing to allow, and some of the very high residuals are less ancient than he considers them, since they are, in reality, parts of younger surfaces greatly upheaved.

Some local multiplicity of surfaces found in various parts of Africa is explained by King as "due to splitting or multiplication of nickpoints during upstream travel of the cycle. Such splitting may be due to a variety of causes: plain zones of resistant rock, zones of hard rock encountered in stenciling through a Karroo cover, later transverse warping, and so forth."

King has mapped "land surfaces due to major erosion cycles" (pl. 2), and remarks that all southern and central Africa may be so "parcelled out." His comments on the lines along which "one cycle impinges on another" are best read with an eye on the Natal map, but may also be quoted here:

> The transition from one cycle to another may be abrupt or gentle, along a line or along a zone. . . . Sometimes the zone of transition is rough

> and hilly, with the hills showing part of the upper level and the valleys joining downstream into the lower level. . . . [A description by H. B. Maufe is quoted in support.] Perched drainage and river captures are numerous near the encroachment of new cycles, and in some cases watersheds have migrated remarkably. . . .
>
> Probably at no time did a single cycle occupy even half of Southern Africa. To think of the whole countryside as planed, say in the Miocene, is probably quite erroneous. While gaining territory in the heart of the continent the Miocene cycle was being destroyed for a considerable distance from the coast. The cycles, growing by pedimentation, chase each other from the coast or up the main drainage lines at fairly constant speed, and the necessary postulate under the peneplanation concept of diminishing intervals between successive uplifts appears not to be necessary under pediplanation. Thus, by co-existence, certain of the cycles can, if that is desired, be regarded as older than has hitherto been the custom.
>
> Even between different drainage basins under the same cycle there is commonly a significant difference of elevation, particularly as between Atlantic and Indian Ocean drainages. . . . The conflict between cycles in contiguous areas has undoubtedly produced migration of divides and change of drainage catchments on a much wider scale than has hitherto been recorded. . . . Competition between the rivers of Southern Africa has been keen, and little doubt can exist that they themselves have changed the form and size of their drainage basins from time to time as one cycle after another swept across the landscape and placed each basin at advantage or disadvantage.

The concept of very gradual development of surfaces by back-wearing of scarps makes it clear that the "actual age," as King calls it, making a distinction from "comparative age, relative to later erosion surfaces," applies

> only to a restricted area, and different parts of the same cyclic surface may have different "actual" ages. A cycle may be current in the interior long after it has terminated near the coast. I should be prepared to learn, for instance, of a cycle initiated at the coast in early Cretaceous time being of late Tertiary actual age in the interior. . . . (This is more or less the history of the "Miocene" or "mid-Tertiary" plain in Africa.)

Again he remarks concerning this "Miocene" surface:

> It was . . . in the peripheral regions late Jurassic or early Cretaceous in age, extending into the interior, spreading by the process of slope retreat from the valley sides of entrenching streams across the whole interior by about the Miocene. . . . In Kavirondo it was deformed even before the Miocene. In the Congo-Angola area it bears silicified gravels and sands of Oligocene age; in Bushmanland it is said to bear late Cretaceous sediments. . . . In the opposite time-sense the same "Miocene" cycle is current in some parts at the present day.
>
> Despite these differences of age from place to place, this great erosion plane, identified in all the territories in Central and Southern Africa, forms the most reliable interior datum for the discrimination of subsequent earth movements and for purposes of correlation. The "late Jurassic" and "late Cretaceous" surfaces described by Dixey, though widespread, are found only as the merest remnants. . . . It was the "Miocene" surface . . . which was the first to sweep as a more or less single feature across the subcontinent.

Quoting figures of rate of retreat of scarps in front of advancing pediplanes, King states:

> I have made two estimates based on different data for the rate of retreat of the Drakensberg; these were 1 foot in 160 years and 1 foot in 233 years. For the Town Hill scarp behind Pietermaritzburg an estimate of retreat is 1 foot in 120 years.

C. A. COTTON

Climatic Accidents in Landscape Making. By C. A. COTTON. 2d printing. New York: John Wiley & Sons, Inc., 1942; printed by Whitcombe & Tombs, Ltd., New Zealand. Pp. xx+354; figs. 149; pls. 58. $7.00.

Professor Cotton's advanced geomorphology dealing with landscapes in dry-climate and glaciated regions is available for the first time from an American publisher. It was reviewed previously in the *Journal*, vol. 51, p. 556, 1943.

L. H.

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ON RATIO CORRELATION IN PETROGRAPHY[1]

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ABSTRACT

The same ratio correlation may be generated by many different combinations of relationships between absolute measures, but a single set of absolute-measure statistics leads to one, and only one, correlation between any particular set of ratios formed from these absolute measures. The passage from ratio correlation to inference about relations between absolute measures is ambiguous at best and often misleading.

Algebraic statements exhibiting ratio correlation as a function of absolute-measure statistics are offered for types of ratios commonly used in petrography. These statements are all derived from Pearson's general formula for index correlation. They yield good approximations only if the fraction $s/\bar{x}$ for each absolute measure is suitably small.

Several practical examples drawn from petrographic literature are described. In most of these cases the ratios seem to have been used either to order the data or in the hope that they would throw some light on relationships between absolute measures. The results are shown to be on the whole indecisive and ambiguous and in a few cases decidedly misleading.

The formation of ratios should be confined to those problems in which hypotheses being tested deal with ratios. Absolute measures are always preferable when large numbers of observations must be recorded without benefit of satisfactory hypothesis. Ratios can always be drawn from tables of absolute measures; frequently, absolute measures cannot be reclaimed from tables of ratios.

INTRODUCTION

In petrography, as in any natural science, one of the principal drives of the observer is to order or "make sense" of his data. We all share the suspicion that the truth, if we could discover it, would be systematic; most of us would probably go even further and confess that when we have discovered a little order in our data we feel that we have come one step closer to the truth. The procedure is logically untenable, for history shows that false systems are possible; but to question its psychological validity would be to question the whole basis of descriptive science. Whether we regard it as a duty or a pleasure, our task is to search for order and even sometimes to invent it when we cannot find it; we cannot help preferring an orderly, systematic result to a disorderly, unsystematic one.

Unfortunately, the search for order easily gets out of hand, and this is particularly the case when it is necessary to record and assimilate large numbers of observations concerning which our hypotheses are either inadequate or indecisive. We do not understand the significance of the observations, but it is still incumbent upon us to record all and remember a good many of them. It is quite as impossible to record a chaos as to remember one, and so we are obliged to

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[1] Manuscript received August 4, 1948.

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impose system where we have not as yet been able to discover it.

Where the observations consist of measurements of two or more variables, some of our commoner notions of order and system are generalized in the concept of correlation. If the measurements are uncorrelated, many ratios formed from them will generally exhibit correlation. Purely for the sake of efficient description, the ratios seem more useful than the absolute values do. Their utility in one situation recommends their application in others, and we tend to overlook the fact that ratio correlations rarely permit unambiguous inferences about absolute values.

Ratio correlation is an old subject in statistics—so old, in fact, that it is not even mentioned in some of the newer texts.[2] The purpose of this paper is to restate, in petrographic context, some of the more important properties of ratios in correlation analysis. No revolutionary advance is likely to result from the attempt, but I hope it will intrigue other petrographers, as it has intrigued me, to discover that what we have learned piecemeal and with much pain has been available for the asking in the journal and text literature of a sister-science for over half a century.

DEFINITIONS AND TERMS

Definitions 1–3 are used to reach definitions 4 and 5; only 4 and 5 are required in the discussion which follows.

1. $\bar{x}_k$ is the mean of $X_{k1}, X_{k2}, X_{k3}, \ldots, X_{kn}$ measurements, or
$$\bar{x}_k = \frac{1}{n}\sum_{i=1}^{n} X_{ki}\,.$$
2. x_{ki} is the deviation of the ith X_k from the mean of the X_k's, or
$$x_{ki} = (X_{ki} - \bar{x}_k)\,.$$
3. s_k is the standard deviation, or square root of the variance, of the X_k's, or
$$s_k = \left(\frac{1}{n-1}\sum_{i=1}^{k} x_{ki}^2\right)^{1/2}.$$
4. C, the coefficient of variation, is the ratio of standard deviation to mean, or
$$C = \frac{s_k}{\bar{x}_k}\,.$$
5. r, the product moment coefficient of correlation, is the ratio of the covariance of two variables to the geometric mean of their variances, or
$$r_{12} = \frac{\frac{1}{n-1}\sum_{i=1}^{n}(x_{1i}x_{2i})}{s_1 s_2} = \frac{p_{12}}{s_1 s_2}\,.$$

The correlation coefficient is a remarkably useful index of the relation between two variables. It will always be true that $-1 \leqslant r \leqslant 1$; if $r = |1|$ the variables stand in perfect linear relation to each other, while if $r = 0$ there is no linear correlation. If the variability be expressed as variance, r^2 gives the proportion of it attributable to linear regression of either variable on the other; $(1 - r^2)$ is the amount of "residual" variation which cannot be described in this fashion. If $r > 0$, the quantities involved vary directly; if $r < 0$ their variation is inverse. We use r in this note purely as a sample description, so that information about the distribution of the parent deviates is unnecessary.

PEARSON'S GENERAL FORMULA FOR RATIO CORRELATION

More than half a century ago, K. Pearson (1896–1897) showed that, if ratios were formed from absolute measures, such that

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[2] It is treated briefly by Snedecor, *Statistical Methods* (4th ed., 1946), pp. 162–64, and at greater length by Yule and Kendall, *An Introduction to the Theory of Statistics* (13th ed., 1946), pp. 299–301.

$$Y=\frac{X_1}{X_2} \quad \text{and} \quad Z=\frac{X_3}{X_4}, \tag{1}$$

the correlation between the ratios was implicit in the variances, covariances, and means of the absolute measures. His derivation assumed that the absolute measures were normally distributed and that in each case the coefficient of variation was small enough that its third and higher powers could be ignored. Reed (1921) later arrived at the same result without specifying the parent-distribution of the absolutes. For our purpose it is important to note that the remaining restriction, governing the size of the C values, is required only to reach a convenient algebraic formulation; correlation between the ratios is a single-valued function of the variances and covariances of the absolute measures, whether or not the relationship may be conveniently stated. Pearson's general formula, which is rarely given in elementary texts, is

$$r_{yz}=\frac{r_{13}C_1C_3-r_{14}C_1C_4-r_{23}C_2C_3+r_{24}C_2C_4}{(C_1^2+C_2^2-2\,r_{12}C_1C_2)^{1/2}\,(C_3^2+C_4^2-2\,r_{34}C_3C_4)^{1/2}}. \tag{2}$$

It contains quantities of only two types, and these are symmetrically distributed. The denominator will always be positive, so the sign of r_{yz} depends on the sign of the numerator. The quantity r_{yz} will be zero only if $r_{13}=r_{14}=r_{23}=r_{24}=0$;[3] if one of these correlations is not zero, r_{yz} will also differ from zero. But it may differ in sign from the nonzero r on the right, and it will almost certainly differ in size. A single combination of r and C values can lead to one and only one value for r_{yz}, but the same r_{yz} might result from many different combinations of r and C values; this is the foundation for the assertion that correlations between ratios do not permit unambiguous inferences about the variables from which the ratios are formed.

SPECIAL CASES OF RATIO CORRELATION

By manipulating the absolute measures it is possible to develop from equation (1) simpler formulas for ratio correlations covering a wide variety of practical problems. In this section the more important relations are listed, and in the succeeding section a few practical applications are described.

CORRELATION BETWEEN TWO RATIOS WITH COMMON DENOMINATOR

If, in equation (1), $X_2 = X_4$, so that

$$Y=\frac{X_1}{X_2} \quad \text{and} \quad Z=\frac{X_3}{X_2}, \tag{3}$$

then $C_2 = C_4$, $r_{24} = 1$, $r_{12} = r_{14}$, $r_{23} = r_{34}$, and equation (2) reduces to

$$r_{yz}=\frac{r_{13}C_1C_3-r_{12}C_1C_2-r_{23}C_2C_3+C_2^2}{(C_1^2+C_2^2-2\,r_{12}C_1C_2)^{1/2}\,(C_3^2+C_2^2-2\,r_{23}C_3C_2)^{1/2}}. \tag{4}$$

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[3] Or if the terms in the numerator cancel completely—a very unlikely circumstance.

This type of ratio correlation is common in many descriptive sciences, for it is often convenient to use one variable as a sort of scaling unit. If the absolute measures are uncorrelated (e.g., $r_{12} = r_{13} = r_{23} = 0$), equation (4) reduces to

$$r_{yz} = \frac{C_2^2}{(C_1^2+C_2^2)^{1/2}(C_3^2+C_2^2)^{1/2}}, \quad (5)$$

which is Pearson's definition of "spurious correlation" and will always be positive. Even in this greatly simplified form, many combinations of C values might yield the same r_{yz}. For the special case in which $C_1 = C_2 = C_3$, $r_{yz} = 0.5$; if C_2 is notably larger than C_1 and C_3, r_{yz} may be much larger than 0.5. If $C_1 = C_3 = C_2/2$, for instance, $r_{yz} = 0.67$. If $C_1 = C_3/2 = C_2/3$, $r_{yz} = 0.79$.

CORRELATION OF A RATIO WITH A THIRD VARIABLE

In equation (1) let $X_4 = 1$, so that

$$Y = \frac{X_1}{X_2} \quad \text{and} \quad Z = X_3 . \quad (6)$$

Then equation (2) reduces to

$$r_{yz} = \frac{r_{13}C_1 - r_{23}C_2}{(C_1^2+C_2^2 - 2\,r_{12}C_1C_2)^{1/2}} . \quad (7)$$

If neither term of the ratio is correlated with Z, $r_{yz} = 0$. If there is positive correlation between X_1 and X_2, the size of r_{yz} is increased, but its sign is unaffected. If $r_{13} = r_{23}$, the sign of r_{yz}, and to a large extent its size as well, will depend on the sizes of C_1 and C_2. If the terms of the ratio are completely uncorrelated ($r_{12} = 0$), r_{yz} will nevertheless differ from zero, provided only that *either* r_{13} *or* r_{23} does so.

CORRELATION OF A RATIO WITH ITS OWN DENOMINATOR

In equation (1) let $X_2 = X_3$ and $X_4 = 1$, so that

$$Y = \frac{X_1}{X_2} \quad \text{and} \quad Z = X_2 . \quad (8)$$

Then equation (2) reduces to

$$r_{yz} = \frac{r_{12}C_1 - C_2}{(C_1^2+C_2^2 - 2\,r_{12}C_1C_2)^{1/2}}, \quad (9)$$

which will be positive only if $r_{12}C_1 > C_2$.

If $r_{12} = 0$, equation (9) reduces to

$$r_{yz} = \frac{-C_2}{(C_1^2+C_2^2)^{1/2}}, \quad (10)$$

so that the "spurious" correlation is here negative. If, further, $C_1 = C_2$, $r_{yz} = -0.71$, and if $C_2 > C_1$ it will be even larger; if C_2 is only twice as large as C_1, for instance, $r_{yz} = -0.89$.

CORRELATION OF A RATIO WITH ITS OWN NUMERATOR

In equation (1) let $X_1 = X_3$ and $X_4 = 1$, so that

$$Y = \frac{X_1}{X_2} \quad \text{and} \quad Z = X_1 . \quad (11)$$

Then equation (2) reduces to

$$r_{yz} = \frac{C_1 - r_{12}C_2}{(C_1^2+C_2^2 - 2\,r_{12}C_1C_2)^{1/2}} . \quad (12)$$

If $r_{12} = 0$, equation (12) further reduces to

$$r_{yz} = \frac{C_1}{(C_1^2+C_2^2)^{1/2}}, \quad (13)$$

the "spurious" correlation again being positive, as in Pearson's type case. If, in addition, $C_1 = C_2$, $r_{12} = 0.71$; if $C_1 > C_2$, it will be larger.

CORRELATION BETWEEN TWO RATIOS WITH THE SAME NUMERATOR

This case is complementary to the original definition of spurious correlation (eq. 5) and is reached from equation (2) in similar fashion. If, in equation (1), $X_1 = X_3$, so that

$$Y = \frac{X_1}{X_2} \quad \text{and} \quad Z = \frac{X_1}{X_4}, \tag{14}$$

equation (2) reduces to

$$r_{yz} = \frac{C_1^2 - r_{14}C_1C_4 - r_{12}C_1C_2 + r_{24}C_2C_4}{(C_1^2 + C_2^2 - 2\,r_{12}C_1C_2)^{1/2}(C_1^2 + C_4^2 - 2\,r_{14}C_1C_4)^{1/2}}. \tag{15}$$

If $r_{12} = r_{24} = r_{34} = 0$, equation (15) reduces to

$$r_{yz} = \frac{C_1^2}{(C_1^2 + C_2^2)^{1/2}(C_1^2 + C_4^2)^{1/2}}. \tag{16}$$

As in equation (5), the spurious correlation here is positive. If, in equation (16), $C_1 = C_2 = C_4$, $r_{yz} = 0.5$, exactly as in equation (5).

For the sake of simplicity the order of subscripts set up in equation (1) has been followed throughout. Notice, however, that equation (16) is actually identical with equation (5). In each case the numerator is the square of the coefficient of variation of the common term, and the denominator is the product of the square roots of the sums of squares of each unique term C with the common term C. A similar term-by-term identity holds for equations (15) and (4). Thus the ratio correlation will be identical whether the common term is used as numerator or as denominator, provided only that it occupies the same position in Y as in Z.

CORRELATION BETWEEN RATIOS WHEN THE DENOMINATOR OF ONE IS THE SAME AS THE NUMERATOR OF THE OTHER

If, in equation (1), $X_2 = X_3$, so that

$$Y = \frac{X_1}{X_2} \quad \text{and} \quad Z = \frac{X_2}{X_4}, \tag{17}$$

equation (2) reduces to

$$r_{yz} = \frac{r_{12}C_1C_2 - r_{14}C_1C_4 + r_{24}C_2C_4 - C_2^2}{(C_1^2 + C_2^2 - 2\,r_{12}C_1C_2)^{1/2}(C_2^2 + C_4^2 - 2\,r_{24}C_2C_4)^{1/2}}; \tag{18}$$

and, if $r_{12} = r_{14} = r_{24} = 0$, equation (18) reduces to

$$r_{yz} = \frac{-C_2^2}{(C_1^2 + C_2^2)^{1/2}(C_2^2 + C_4^2)^{1/2}}. \tag{19}$$

If, in addition, $C_1 = C_2 = C_4$, $r_{yz} < -0.5$; if C_2 is greater than C_1 and C_4, $r_{yz} < -0.5$. Thus if the absolute measures are uncorrelated, the ratio correlation in this case will be equal in size and opposite in sign to that between ratios with common denominator or common numerator. In the same way, equation (18) will yield correlations equal in size and opposite in sign to those given by equations (15) and (4).

SOME PETROGRAPHIC RATIO CORRELATIONS

In this section some practical instances of petrographic ratio correlation are reviewed. The use of ratios is very common in petrography, and, although we often neglect to compute the actual correlation, we almost invariably arrange the paired values in tables or graphs which accept all the hazards but share few of the advantages of results obtained by computation. The examples cited here are of uneven grade and have in common only the property that each has come to my attention during the course of the past few months. In some cases computations have been carried out, in others not. In some instances the usual interpretation of results passes from correlations of ratios to inferences about absolute measures, while in others due allowance is normally made for the algebraic consequences of forming the ratios.

THE GENERAL CASE: TWO RATIOS WITHOUT COMMON TERM

G. T. Prior suggested in 1916 that in the stony meteorites the ratio MgO/Nonmetallic Fe was negatively correlated with the ratio Ni/Metallic Fe. He seems to have considered this negative ratio correlation evidence for his hypothesis that the oxidized iron of stony meteorites formed at the expense of metallic iron. If the oxidized iron formed in this fashion, there would under most circumstances be a negative correlation between oxidized and metallic iron; hence I think it is fair to say that Prior's argument passes directly from a correlation between ratios to an inference about relations between two of the absolute measures from which the ratios are formed.[4] The absence of a common term renders the case rather complex, but factors influencing the sign of the ratio correlation may be visualized by entering equation (2) with X_1 = MgO, X_2 = Nonmetallic Fe, X_3 = Ni, X_4 = Metallic Fe. From the fact that r_{yz} is negative Prior infers that r_{24} is also negative. From equation (2) it is clear that r_{24} is only one of four terms (containing combinations of eight variables) governing the sign of r_{yz}. Now there is, in fact, a very high negative correlation between the two forms of iron ($r_{24} = -0.68$ as computed from the table given by Brown and Patterson [1947*a*]), but equation (2) shows that r_{yz} might be negative *regardless* of the value of r_{24}. Only if the sum of the first three terms was equal to or larger than zero could it be argued that $r_{yz} < 0$ required $r_{24} < 0$.

The passage from knowledge about r_{yz} to inference about r_{24} would be on somewhat firmer ground if we had reason for supposing that the sum of these first three terms was not negative; but the simplest unexceptionable way to determine the relation between X_2 and X_4 is to compute r_{24}. In a situation as complex as this, the sign of the ratio correlation is not a safe index of the sign of any one of the absolute-measure correlations.

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[4] Perhaps it would be even fairer to say that what I am criticizing is my own—and I think the common—impression of Prior's position. In 1913 he argues (p. 35) that except for variations in the amount of nickel-iron, chondrites are, on the whole, characterized by "remarkable similarity, if not specific identity" in chemical and mineral composition. In 1916 (p. 26) he points out that ". . . the less the amount of nickel-iron in chondritic stones, the richer it is in nickel and the richer in iron are the magnesium silicates"; but in his classification ". . . the main principle . . . is the ratio of MgO to FeO in the magnesium silicates and in its interrelation with the richness in nickel of the nickel-iron." In 1919 (p. 350) he clearly restates his belief in the oxidation hypothesis as an explanation of the dual ratio correlation (which in the language of this paper amounts to inferring a relation between absolutes from a relation between ratios), regards the actual amounts of iron only as "support to this view," and remarks that the amount of nickel in the metallic portion of meteorites is ". . . the factor which is seen to determine many of their other characters."

Another instance of dual ratio correlation is Washington's well-known comparison of

$$\frac{K_2O}{K_2O+Na_2O} \text{ with } \frac{MgO}{MgO+FeO}$$

in analyses of igneous rocks (1915). Figure 1 is his graph of the relationship, and from it he argues that ". . . in igneous magmas potassium and magnesium on the one hand, and sodium and iron on the other, tend to vary together." This conclusion is open to a variety of interpretations. The first impression of most people seems to be that it asserts the existence of two positive correlations, one between potash and magnesia, one between soda and iron. Some readers feel that the correlations but not their signs are specified. After a little consideration this interpretation is commonly dropped in favor of one suggesting that in "potash-rich rocks" the "magnesia ratio" is usually high, and conversely. For this purpose "potash-rich" may mean either that the ratio of potash to soda is large, that the ratio of potash to the sum of soda and potash is large, or just that the rocks are high in potash; and "magnesia ratio" similarly may mean either the ratio of MgO to FeO or the ratio of MgO to the sum of MgO plus FeO. Washington's own interpretation conforms to the first impression described above, viz., ". . . soda not uncommonly tends to vary with the iron oxides, while potash shows similar relation to magnesia." The correlation of these two ratios, however, is a function of at least ten absolute-measure correlation coefficients, only four of which are not potentially quite independent of one

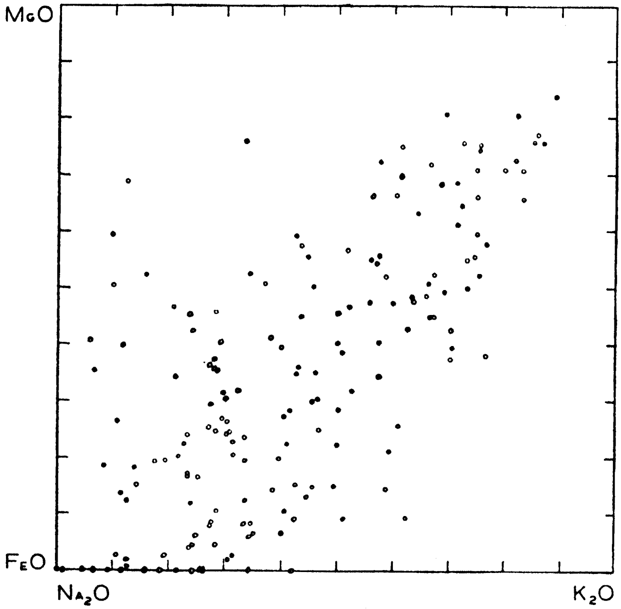

FIG. 1.—The ratio of MgO to the sum of MgO and FeO as a function of the ratio of Na_2O to the sum of Na_2O and K_2O in analyses of igneous rocks (from Washington).

another. It is certainly unreasonable to expect that valid inferences about relations between any particular pair of absolute measures may be deduced from the mere fact that the ratio correlation is positive. If the purpose is to determine whether or not potash and magnesia are positively correlated, their correlation should be computed directly from the data. If it is desired to check the effect of soda, iron, or some function of the two on this relationship, partial correlation should be used.

In any case, no one of the interpretations advanced here may be regarded as favored by figure 1. The ratio correlation means exactly what it says—that the ratio of potash to the sum of potash and soda varies directly with the ratio of magnesia to the sum of iron and magnesia. The graph suggests that, where sizable amounts of all four constituents are present, the correlation is, in fact, very high, and the relationship may be of considerable use if it is supplemented by information about relations between absolute measures; such information, however, may not be inferred from the ratio correlation but must be obtained directly from the data.

TABLE 1

VALUES OF r_{yz} FOR 14 ANALYSES OF APATITE IRON ORES FROM NORTHERN SWEDEN

Y	Z			
	V_m/V_n	Mn_m/Mn_n	Co_m/Co_n	Ni_m/Ni_n
P_2O_5/SiO_2...	+0.04	0.71	0.80	0.78
P_2O_5......	−0.04	0.68	0.84	0.85

As a final example of this case some calculations recently given by Landergren (1948) are shown in table 1. In the table the subscript *m* means "magnetic," *n* means "nonmagnetic." These results are useful for two reasons. They show the utility of the general relations, provided that the restriction on the size of the *C* values is observed, and they present an opportunity for suggesting that this condition will sometimes be violated in practice. Disposing of this latter point first, minor constituents may vary widely in individual analyses, but, by definition, their averages are small. Therefore, *C* for a minor constituent may often be large enough that its third and higher powers cannot be ignored. The approximations given in the preceding section will therefore be unsatisfactory.

On the assumption that *C* values are small enough here that equation (4) is still applicable, it may be suggested that Landergren's interpretation of table 1 is open to question. He concludes that ". . . the quantity of Mn, Co and Ni entering the magnetic fraction and the nonmagnetic fraction respectively is dependent on the . . . content of P and the ratio P_2O_5/SiO_2." Now the correlations in the first line of table 1 conform exactly to equation (2) of the preceding section, while those of the second line fall under the case described by equation (7). Agreement between the two lines of the table is very good, so that one is led at once to wonder under what circumstances equations (2) and (7) might lead to similar results. Equation (7) is reached from equation (2) by deleting all terms involving X_4; table 1 would be explained if all correlations involving SiO_2 were negligibly small and the coefficient of variation of silica were considerably smaller than that of phosphorous pentoxide.

It thus seems quite possible that silica is not sensibly correlated with *any* of the absolute values, and it is hard to see why the various *m*/*n* ratios of table 1 should be regarded as functions of silica unless this can be established directly from the data. Unfortunately, Landergren has not recorded the basic measurements, and some or all of the C values may well be high enough to invalidate arguments based on equations (2) and (7); the case is included as an illustration of the power of the method under more auspicious circumstances.

CORRELATION BETWEEN TWO FRACTIONS WITH COMMON DENOMINATOR

The relation which first attracted Pearson's attention and on which he based the definition of spurious correlation was the correlation between two fractions with common denominator. In it fall most of the major parameters of the continental chemical-petrographic systems. As a hypothetical example, suppose that the quantities of olivine, pyroxene, and garnet are uncorrelated in a series of contact marbles. Letting olivine be X_1, pyroxene X_2, garnet X_3, we have $r_{12} = r_{13} = r_{23} = 0$. If two ratios, Y and Z, are formed from each set of measurements by using any one variable as denominator over the entire group, then $r_{yz} > 0$, *regardless of which variable is used as denominator*. If $C_1 = C_2 = C_3$, $r_{yz} = 0.5$. If the object is to obtain as high a correlation as possible, the variable with largest C, often the one with smallest mean, should be used as common term.

Incidentally, comparison of equations (5) and (16) indicates that this rule holds, regardless of whether the common term is numerator or denominator. Further, comparison of equation (4) with equation (15) shows that the nonspurious ratio correlation (e.g., that holding when any or all of the absolute-measure correlations differ from zero) will also be identical ***whether the common term is used as numerator or as denominator*** (see discussion following eq. [16]).

The Niggli system affords excellent examples of ratio correlation, for each of its major parameters is the quotient of a molecular amount divided by the sum of several molecular amounts. It is not infrequently supposed that the relations between these parameters give a fair approximation of the relation between molecular amounts, molecular proportions, or weight percentages of the oxides after which they are named.

Figures 2 and 3 illustrate a practical example. In each figure a graph of the Niggli symbols *alk* and *qz* for a group of analyses is shown on the left, and a graph of normative (*or*+*ab*) and normative Q for the same analyses appears on the right. Normative *or* and *ab* may be regarded as multiples of the molecular amounts or weight percentages of the alkalis, whereas normative Q is an estimate of the molecular amount or weight percentage of uncombined silica. Relations between the Niggli symbols are very different from those exhibited by the normative parameters. I have shown elsewhere (1948) that some approximation of the relation between modal quartz and alkali-feldspar is just about the most important information to be extracted from these analyses; evidently, the result obtained will depend on the system used for reduction of the chemical analyses. In this particular case the Niggli parameters are misleading in the extreme, though the correlations between Niggli symbols are quite as "real" and "true" as the correlations between normative molecules.

CORRELATION OF A RATIO WITH A THIRD VARIABLE

The correlation of a ratio with a third variable is not a particularly striking case, for the ratio correlation will be zero if neither term of the ratio is correlated with the third variable. It is to be noted from equation (7) that the coefficient of variation of the third variable does not directly affect the correlation between it and the ratio. The connection between equations (2) and (7) has already been described, with Landergren's computations as an example.

If one is simply attempting to systematize unruly data by the use of this type of ratio correlation, equation (7) shows that, in general, the highest ratio correlation will be found if the ratio is formed from the two variables exhibiting the strongest direct variation.

Nockolds (1947) has recently suggested that in aplites the ratio *or*/(*or* + *ab*) varies directly with normative quartz, and the same conclusion was reached by Vogt (1930) with regard to granites some years ago. One would suppose that in a case as simple as this it would be possible to pass directly from ratio correlation to inferences about

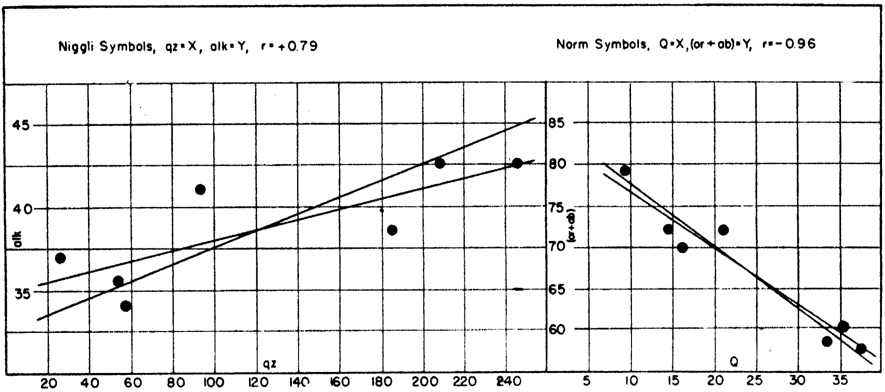

FIG. 2.—Niggli and Norm symbols, showing relation between alkalis and uncombined silica, rapakivi granites of Loos Hamra (data from H. von Eckermann, Förh. Geol. Fören., vol. 58, no. 405, 1936).

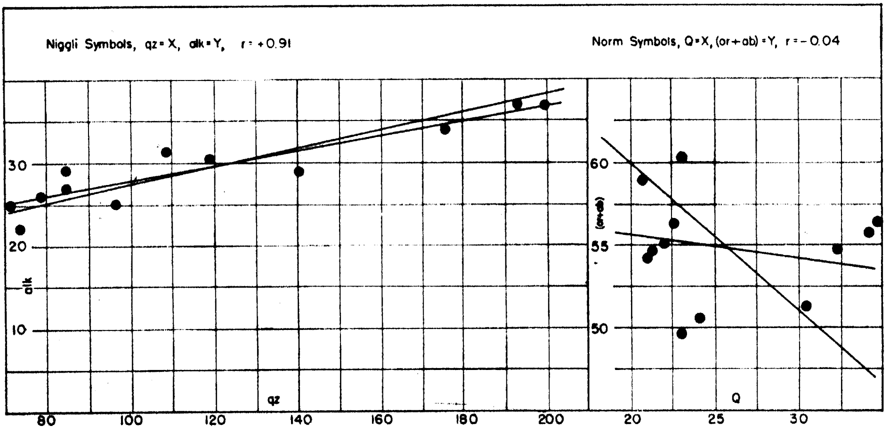

FIG. 3.—Niggli and Norm symbols showing relation between alkalis and uncombined silica, rapakivi granites of Lappeenranta (data from V. Hackman, Comm. géol. Finlande Bull. 106, 1934).

absolute measures. Equation (7) indicates that this is not so; for the ratio correlation to be positive, it is necessary only that $r_{13}C_1 - r_{23}C_2 > 0$. Both r_{13} and r_{23} may be positive or negative. If $r_{yz} > 0$, we may not argue, for instance, that a highly siliceous granite will, in general, be rich in *or* or that a granite low in silica will, in general, be rich in *ab*+*an*. I ought to add that neither Vogt nor Nockolds makes such inferences from the mere fact of the ratio correlation. Many of us, however, do make statements of the sort, usually based on direct knowledge of the absolute measures; and in unwary moments nearly all of us imagine that the ratio correlation in some way confirms or even controls the observed relations between the absolute measures.

In a comprehensive study of the Odenwald diorites, Erdmannsdörffer has used a ratio correlation of this type in an attempt to discover the relation between hornblende, biotite, and quartz. The observed quantities of these three minerals vary widely and irregularly in over fifty modal analyses, but Erdmannsdörffer points out that the ratio *hbl/bi* decreases with increasing quartz. No computations are given, but a graph of the data leaves no doubt that a significant and probably large proportion of the variability of the ratio may be described as a linear function of quartz content. From this, Erdmannsdörffer concludes that "Man erkennt deutlich, wie mit steigendem Quarzgehalt, die Menge des Biotits zunimmt." This should not be regarded as in any sense a necessary consequence of the ratio correlation, though it is, of course, one of many possible explanations. That it is not the correct explanation is suggested by the fact that Erdmannsdörffer was led to use ratios in the first place by the ". . . scheinbare Regellosigkeit . . ." of the modal analyses.

CORRELATION OF A RATIO WITH ITS OWN DENOMINATOR

A common case in petrography is the correlation of a ratio with its own denominator; and our interpretations of it are often correct for the reason that in making them we almost always rely heavily on information about the absolute measures. For some curious reason this information, which is the key to the situation, rarely receives the same prominence as does the ratio correlation.

In his excellent study of the Tasmanian dolerites, A. B. Edwards (1942) has recently described the negative correlation between MgO and the ratio FeO/MgO. It is by now evident that this correlation tells nothing of the relation between FeO and MgO. Equation (10) shows that, even if there is no correlation between FeO and MgO, there will be negative correlation between MgO and the ratio FeO/MgO and that, if the coefficient of variation of MgO is equal to or larger than that of FeO, the absolute value of this correlation will be equal to or greater than 0.71, MgO and FeO still being uncorrelated. This seems to be the case with Edwards' analyses. In each of the sills studied, MgO has a very much wider range and a somewhat smaller average value than FeO. The ratio correlation is thus compatible with the notion that there is no systematic relation whatever between MgO and FeO.

What gives meaning to Edwards' data is the persistence with which high MgO values occur in the lower, and low MgO values in the upper, portions of each sill. MgO has surely been extracted from the liquid during crystallization, but with little if any effect on the FeO content of the residue. One of the unfortunate by-products of our fascination with ratios is that we pay insufficient attention to the absolute measures. The most prominent feature of one of Edwards' oxide profiles is the sharp increase in alumina, matching an equally sharp decrease in MgO, but apparently unrelated to silica; and throughout all the analyses there seems to be an inverse variation of alumina and magnesia. Yet the interesting possibility that alumina proxies for magnesia in the pyroxenes rather early in the crystallization is not considered. Something of the sort would seem to be almost inevitable unless color index increases considerably and systematically with depth.

Another example of this case is the correlation of Ni/Total Metal with Total Metal in stony meteorites. First discovered by Flight (1887), this relationship appears again as an introduction to the dual ratio correlation proposed by Prior and has recently been revived by Brown and Patterson (1947*a*, *b*; 1948), to whom we are indebted for the first reasonably complete compilation of stony meteorite analyses in many years. The earlier work was based on rather few analyses, but the Brown-Patterson table contains fifty-three in the critical range and makes it possible to establish the relationship as something more than a shrewd guess. These authors have computed average Ni/Metal values for 2 per cent intervals of metal content and have plotted the average values of the ratio as a function of its denominator. Their graph of the relationship is shown here as figure 4. As most of the metal content is iron, one would expect a similar relation to hold between Ni/Metallic Fe and

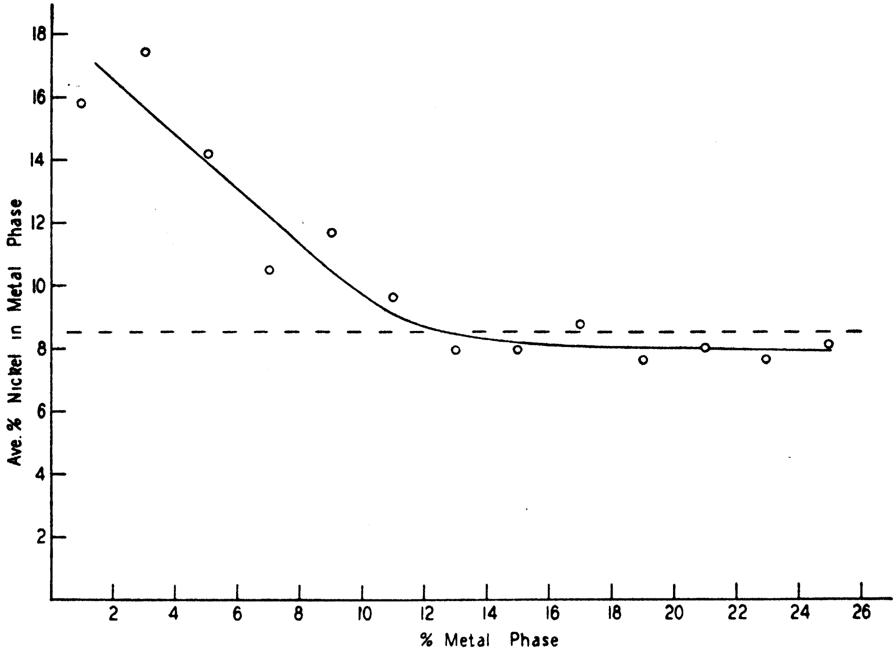

FIG. 4.—The ratio $\frac{\text{Metallic Ni}}{\text{Total Metal}}$ as a function of total metal in stony meteorites (from Brown and Patterson).

Metallic Fe, and this proves to be the case.

For the analyses listed in the Brown-Patterson table (1947*a*, p. 405), the coefficient of variation of Ni is 0.49, and that of Fe is 0.60; under the circumstances equation (9) shows that nothing can prevent the occurrence of negative correlation between Ni/Metallic Fe and Metallic Fe; for practical purposes the range of this ratio correlation is $0 \geqslant r_{yz} \geqslant -1$.[5]

In forming ratios we are always careful to use quantities occurring in the same analysis; in forming, say, the eighth ratio of a group, we divide the Ni value reported in the eighth analysis by the metallic Fe value in the same analysis, and the comparison is between the resulting fraction and its denominator. What equation (10) tells us is that this precaution is quite unnecessary. The result would be the same, to the required precision, if we just as carefully *avoided* using the true sample pairs. As an illustration, I purposely randomized the Fe and Ni values in all the Brown-Patterson analyses containing less than 14 per cent Metallic Fe[6] and repeated the averaging calculations previously carried through with the true sample pairs. The results are shown in table 2. The tendency for the average ratio value to decrease with increase in Metallic Fe is quite as strong in the randomized pairs as in the true sample pairs. The explanation is not far to seek. The dispersion of Ni values is very much smaller than the dispersion of Metallic Fe. A little consideration will show that, when the range of the numerator is so much smaller than that of the denominator, the fraction is bound to decrease with increase in the denominator, regardless of the relation between numerator and denominator. As a corollary, we ought not to expect to discover much about this relationship from the ratio correlation. As a matter of fact there is very strong positive correlation between Ni and Metallic Fe in the analyses listed by Brown and Patterson.

TABLE 2

AVERAGE NI/METALLIC FE VALUES FOR 2 PER CENT INTERVALS OF FE
TRUE AND RANDOM PAIRS

	MID-POINT OF FE CLASS						
	1	3	5	7	9	11	13
True sample pairs	0.122	0.136	0.122	0.116	0.090	0.097	0.086
Randomized pairs	0.689	0.167	0.183	0.126	0.101	0.072	0.056

CORRELATION BETWEEN TWO RATIOS WITH COMMON NUMERATOR

It has been shown that correlations between two ratios with common numerator will be the same as those found when the common term is the denominator; the discussion under "correlation between two fractions with common denominator" of the present section applies directly to this case.

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[5] A C value of 0.5 is really too high to permit rigorous application of the approximations given in the preceding section; C^3 is 0.125, which may not be neglected if one wants an r value good even to the first digit. The demand for negative correlation is in this case so extreme, however, that the general outline of the argument is applicable, though the exact value of the correlation no doubt differs from that yielded by eq. (9).

[6] By numbering the Ni values in order, assigning to each Metallic Fe value one of the first fifty-three nonidentical two-digit numbers drawn from a page of Tippett's table of random numbers, ranking the results, and combining each Ni value of order i with the Metallic Fe of rank i, where $i = 1 - 53$, inclusive.

CORRELATION OF A RATIO WITH ITS NUMERATOR

Correlations of a ratio with its numerator will be identical in size with those between ratio and denominator, but they will be of opposite sign. Between equations (12) and (9) there is the same parallelism as has previously been noted between equations (18) and (5) (see discussion following eq. [19]).

Correlation of a ratio with its numerator is quite commonly used in geochemical work, as, for instance, in the comparison of Li/Mg with Li by Strock (1936) or of Ge/Si with Ge by Goldschmidt and Peters (1933). Even if the ratio correlation were high, which is usually far from the case, it would not justify a conclusion that there was or was not some systematic relationship between the absolute measures. If the absolute measures are uncorrelated, the ratio correlations will always be positive unless the numerator is constant. In geochemical work, particularly on minor elements, dispersion will often be large in relation to mean, so that the algebraic approximations given here will be unsatisfactory. This does not mean that the ratio correlation is independent of the absolute-measure statistics and is of course no reason for supposing that the latter are determined by the former. For the most part, satisfactory interpretations of ratio correlations will still hinge on adequate information about relations between the variables from which the ratios are formed.

CORRELATION OF TWO RATIOS WHEN THE DENOMINATOR OF ONE IS THE SAME AS THE NUMERATOR OF THE OTHER

With regard to location of the common term, equation (18) is just equation (16) or equation (5) with all signs changed. Thus, if $Y = qz$ of the Niggli system and $Z = alk$, $r_{yz} = -r_{y\frac{1}{z}} = -r_{\frac{1}{y}z}$. If one were studying the relations between MgO, Metallic Fe, and Nonmetallic Fe in meteorites by means of ratios, he would find that this equality also held where, for instance,

$$Y = \frac{\mathrm{MgO}}{\mathrm{Fe}_m}, \qquad Z = \frac{\mathrm{Fe}_n}{\mathrm{Fe}_m}.$$

SUMMARY AND CONCLUSIONS

When four variables are formed into two ratios, there will be no correlation between the ratios if there is no correlation between the variables, and there will, in general, be correlation between the ratios where there is correlation between one or more pairs of the absolute measures if these pairs include a term from each ratio. But the size of the ratio correlation will, in general, differ from that of any one of the absolute-measure correlations, and its sign will provide no indication about the sign of any one of these.

When three variables are formed into two ratios by using one variable as a common term, the size of the correlation will be the same whether the common term appears as the numerator in both ratios, as the denominator in both ratios, or as numerator in one and denominator in the other. In the first two cases the sign of the correlation will also be the same, but in the third it will be reversed.

When three variables are compared by forming any two into a ratio, the correlation of this ratio with the third variable can be zero only if each term of the ratio is uncorrelated with the third measure or if the terms of the ratio are merely coded values of each other. The ratio correlation is not a reliable criterion of the absolute-measure correlations.

When a ratio is formed of two absolute measures and compared with one of the two, the sign of the correlation will depend on whether the common term appears as numerator or as denominator, but its numerical value will be the same in either case.

Whenever there is a common term involved, the ratio correlation cannot be zero if the absolute measures are uncorrelated. In each specific case its sign is fixed; in all cases its size is determined by the relative variability of the absolute measures.

There are very few exceptions to the general rule that no ratio correlation permits valid inference about the relationship between any two of the absolute measures from which the ratios are formed. On the other hand, a knowledge of the sample statistics of the absolute measures will permit valid deduction of the correlation between any set of ratios formed from them, provided only that the *C* values involved are sufficiently small.

In descriptive petrography we commonly resort to ratios when the absolute measures seem feebly correlated or uncorrelated, and we frequently employ one or another of the common-term relationships; so it is not surprising that we have turned up a number of remarkably high ratio correlations. A good deal of the significance that we attach to these correlations stems from the premise that in other samples or in other arrangements the same absolute measures, still uncorrelated, might yield very much smaller ratio correlations or even correlations of opposite sign. On occasion, most of us are not at all averse to passing from correlation between ratios to inferences about absolute measures. On all three counts we often lead ourselves, as well as others, astray.

The misuse of ratios in specific problems is easily corrected. The difficulties encountered in deciding how to record large numbers of measurements are not so readily eliminated; for it may be quite plausibly argued that the justification of the ratio is here purely that of a bookkeeping device. Thus, in at least two instances (Niggli, 1923; Simonen, 1948) the Niggli symbols of large numbers of rock analyses have been recorded in place of the weight percentages or molar amounts. To recapture from these tables enough information for a study of the relationship between any pair of constituents would require almost as much work as the compilation of the table, and for some constituents (e.g., ferrous and ferric oxides) it would be impossible. Similarly, one can discover the composition of any stony meteorite in the Brown-Patterson table, but only by undoing a computation previously performed by Brown and Patterson.

It is very easy to carry an argument of this sort too far. Sometimes hypotheses are framed in terms of ratios, and some phenomena are understandable only as ratios. I do not mean to offer a blanket condemnation of all ratio correlation, but it does seem to me that in most of the problems of descriptive petrography our primary concern is still with the simplest variables obtained by direct measurement or observation. There may be instances in which ratio correlations accomplish more for us than absolute-measure correlations do, but this will usually occur only where our knowledge of the relations between the absolute measures is satisfactory. In any case, the ratio correlations are contained in the absolute measurements, whereas the converse is not usually true. Even when attention is focused on ratios, the absolute measures ought to be recorded, and certainly for purposes of tabulation the absolute measures are preferable.

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SOME EXAMPLES OF THE APPLICATION OF THERMOCHEMISTRY TO PETROLOGY[1]

TH. G. SAHAMA[2] AND D. R. TORGESON[3]

ABSTRACT

A brief summary is presented of measurements of heats of solution of minerals belonging to the forsterite-fayalite and to the enstatite-orthoferrosilite series and of artificial ilmenite and geikielite.

In the olivine and orthopyroxene series, the heat of solution in HF is found to be a linear function of the Mg:Fe ratio, indicating perfect isomorphism between the corresponding end-members. The importance and possibilities of applying calorimetry to the study of isomorphism are emphasized. On the basis of the calorimetric data available for the minerals in question, the heat and free energy are given for the reaction: olivine + quartz → 2 pyroxene. The difference of the stabilities of $MgSiO_3$ and $FeSiO_3$ is illustrated.

The stability reaction of ilmenite and geikielite in the presence of olivine or orthopyroxene is calculated from the calorimetric data. The influence of the entropy of mixing isomorphic minerals upon the stability relation is emphasized.

INTRODUCTION

Besides the direct petrological observations on natural rocks and minerals and experimental work on synthetic compounds, there are still other ways of investigating the formation and stability relations of minerals. One of these methods, indirect in itself, is offered by thermochemistry. Despite the fact that quantitative thermochemistry, based on numerical data, apparently has a wide field of possible application to petrology, it has been used very little as yet for solving petrological and mineralogical problems. The reason for such a circumstance is revealed by the fact that calorimetric data for natural or artificial minerals are scanty. Therefore, it is to be hoped that, in the future, more calorimetric measurements on minerals will be undertaken by scientists interested in problems dealing with the stability relations of minerals and mineral parageneses. A successful application of the thermodynamic principles in petrology evidently is not possible without having available reasonably comprehensive data obtained by experimental calorimetry and covering all the more important rock-forming minerals.

From the viewpoint of the petrologist, the most important thermal quantities for minerals are, perhaps, the heats of formation, owing to the fact that, in a large number of reactions possible between minerals, the largest contributor to the free energy usually is the heat of the reaction. The most suitable method of ascertaining heats of formation of silicates and of other rock-forming minerals is offered by solution calorimetry. This method is applicable to all minerals that are rapidly soluble in hydrofluoric acid or other solvents. If this method is used at elevated temperatures—say, near 75° C.—a considerable number of petrologically important minerals can be investigated.

By means of hydrofluoric acid solution

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calorimetry, the silicates of the olivine and orthopyroxene series were investigated. Previously, Mulert (1912) had found that pure magnesium ortho- and metasilicate did not dissolve rapidly enough to allow a calorimetric heat-of-solution measurement. In the present investigation it was found that, if the material is very finely powdered and if the experiment is made at an elevated temperature, the rate of the reaction can be increased sufficiently for an accurate determination. The determinations were made at 73°.7 C. In addition, some heat-of-solution measurements of ilmenite and geikielite were made. The design of the calorimeter and the procedure employed has been given previously by Torgeson and Sahama (1948). The details of the results have also been reported (Sahama and Torgeson, 1948*a* and *b*). However, it seems pertinent to give in this periodical a brief summary of the mineralogical and petrological results of the work and to point out some possible applications of thermochemical methods and considerations to problems of petrology and mineralogy.

FORSTERITE-FAYALITE SERIES

The samples of the minerals belonging to the forsterite-fayalite series, used for the present investigation, are listed in table 1. In addition, the molecular compositions, calculated on the basis of the chemical analyses as pure mixtures of Mg_2SiO_4 and Fe_2SiO_4, and the measured heats of solution in 20.1 per cent HF (averages of the individual runs), expressed in calories per mol at 73°.7 C., are given in the second and third columns, respectively. The fourth column indicates the uncertainty of the average. All heat-of-solution values are corrected for

TABLE 1

HEATS OF SOLUTION OF THE FORSTERITE-FAYALITE SERIES IN 20.1 PER CENT HF AT 73°.7 C.

Sample	Molecular Composition	ΔH* (Cal/Mol)	Uncertainty (Cal/Mol)
A. Forsterite, Dreis, Eifel, Germany	Fo 90.6 Fa 9.4	−94,250	160
B. Forsterite, Ultenthal, Tyrol, Austria	Fo 90.2 Fa 9.8	−93,590	80
C. Chrysolite, Marjalahti, Finland	Fo 88.2 Fa 11.8	−93,930	140
D. Hyalosiderite, Lake Leistilänjärvi, Nakkila, Finland	Fo 56.0 Fa 44.0	−89,070	100
E. Hyalosiderite, Susimäki, Vampula, Finland	Fo 54.2 Fa 45.8	−88,820	160
F. Ferrohortonolite, Tunaberg, Sweden	Fo 5.9 Fa 94.1	−82,080	160
Fa. Synthetic fayalite	Fo 0.0 Fa 100.0	−81,330	50

* Throughout this paper the symbol ΔH corresponds with the Lewis and Randall usage.

the impurities present in the samples. In designating the natural minerals, the nomenclature proposed by Deer and Wager (1939) was employed.

The results of the heat-of-solution measurements listed in table 1 are plotted against the composition in the upper part of figure 1. As is seen from the figure, the heat of solution of the forsterite-fayalite series is, within the experimental errors, a linear function of the Mg:Fe ratio. The dashed line indicated in the

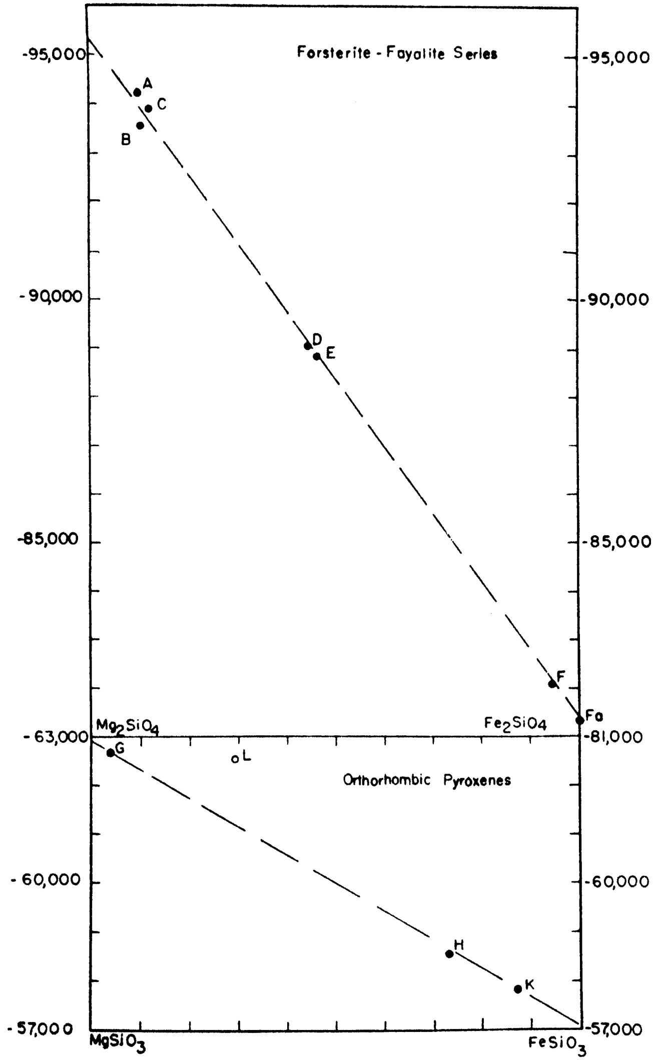

FIG. 1.—Heats of solution in hydrofluoric acid of iron and magnesium silicates: *top*, Mg_2SiO_4-Fe_2SiO_4 series; *bottom*, $MgSiO_3$-$FeSiO_3$ series.

figure gives the best average and can be expressed by the following equation:

$$\Delta H = -81{,}330 - 140.5 X_{mg} ,$$

in which X_{mg} is the molal percentage of Mg_2SiO_4 in the corresponding isomorphic mineral. An extrapolation to pure Mg_2SiO_4 yields

$$\Delta H_{Mg_2SiO_4} = -95{,}380 \text{ cal/mol} .$$

The linearity of the relationship between the heat of solution and the Mg:Fe ratio in the forsterite-fayalite series is mineralogically interesting. It indicates that the replacement Mg-Fe in the olivine structure is thermally perfect. In other words, the heat of isomorphic mixing is zero in this series. Structurally, it means that the isomorphic substitution Mg-Fe in the olivine structure occurs without causing strain.

The structural interpretation of the results of the heat-of-solution measurements expressed above contains in itself nothing new. All petrological observations of natural rocks, supported by the experimental investigations carried out by Bowen and Schairer (1935), indicate that the isomorphism between forsterite and fayalite is complete without any detectable complications. The importance of the result of the heat-of-solution measurements is that the perfectness of the isomorphism can be tested by calorimetric means. It shows the possibility of applying calorimetry to the study of isomorphism of natural silicates and minerals. It is to be expected that many other isomorphic series among the natural minerals will show deviations from linearity, indicating changes in the structure of the mixture as compared with the pure end-members.

THE ENSTATITE-ORTHOFERROSILITE SERIES

The samples of the enstatite-orthoferrosilite series employed, with molecular compositions, heats of solution in 20.1 per cent HF (averages of the individual runs), and uncertainties of these averages are given in table 2. The values are corrected for the impurities. The nomenclature follows Poldevaart (1947).

TABLE 2

HEATS OF SOLUTION OF THE ENSTATITE-ORTHOFERROSILITE SERIES IN 20.1 PER CENT HF AT 73°.7 C.

Sample	Molecular Composition	ΔH (Cal/Mol)	Uncertainty (Cal/Mol)
G. Enstatite, Kjörstad, Bamle, Norway	En 96.5 Fs 3.5	−62,670	80
H. Eulite, Hirvisaari, Rantasalmi, Finland	En 27.1 Fs 72.9	−58,570	50
K. Eulite, Tunaberg, Sweden	En 12.9 Fs 87.1	−57,800	20

The heat-of-solution values given in table 2 are plotted against composition in the lower part of figure 1. The three points for the enstatite-orthoferrosilite series in the figure seem to indicate that in the orthorhombic pyroxene structure also the replacement Mg-Fe is perfect. The average shown by the dashed line in the figure is represented by the equation

$$\Delta H = -57{,}010 - 58.7 X_{mg} ,$$

in which X_{mg} is the molal percentage of $MgSiO_3$ in the mixture. An extrapolation to the pure end-members yields

$$\Delta H_{MgSiO_3} = -62{,}880 \text{ cal/mol},$$

$$\Delta H_{FeSiO_3} = -57{,}010 \text{ cal/mol}.$$

In addition to the samples listed in table 2, a bronzite from Eerolamminvuori, Paistjärvi, Heinola, Finland, was investigated (sample L). Corresponding to its mode of occurrence as a segregation in a labradorite rock, its alumina content is, according to the analysis published by Lokka (1943), unusually high (6.39 per cent Al_2O_3). Its composition is thus outside the enstatite-orthoferrosilite series proper. The mineral shows En:Fs = 70.8:29.2 and gives a heat of solution (20.1 per cent HF at 73°.7 C.) of $\Delta H = -62{,}560 \pm 90$ cal/mol. This value is 1,390 calories above the dashed line in figure 1.

The deviation of the heat of solution of the Heinola bronzite from the heat-of-solution curve of the pure $MgSiO_3$-$FeSiO_3$ series can be qualitatively explained. As pointed out by Lokka (in his paper cited above), aluminum apparently replaces both quadrivalent silicon and divalent iron and magnesium. Now, if trivalent aluminum enters into the structure in place of quadrivalent silicon and divalent iron and magnesium, the electrostatic neutrality will, in general, be disturbed. Even if the structure as a whole remains neutral, individual oxygen ions might not all be equally balanced. The deviation from strict neutrality of a structure is a phenomenon that is very well known in some other minerals (sphene, andalusite, sillimanite, etc.) of comparatively simple chemical composition. In minerals and mineral groups (amphiboles, pyroxenes, micas, etc.) in which the isomorphic substitutions are highly complicated, considerable deviation from neutrality might occur. Qualitatively, it is well understood that an overbalancing or underbalancing of the individual oxygen atoms weakens the structure. Of course, the behavior of the Heinola bronzite represents just one single example of such a supposed weakening of the structure causing a high heat of solution. Therefore, too much weight should not be assigned to it. However, it shows in principle the possibility of applying calorimetric methods to the study of the energy balances connected with the isomorphic substitutions in minerals.

STABILITY OF ORTHOFERROSILITE

As is well known through the investigations carried out by Bowen and Schairer (1932, 1935), pure orthoferrosilite is not stable at moderate and at high temperatures. In the lithophysae of some obsidians from different localities Bowen (1935) has found monoclinic pyroxenes with optical properties corresponding to those extrapolated for pure $FeSiO_3$. However, as pointed out by him, the natural occurrence of ferrosilite in orthorhombic or monoclinic form cannot yet be regarded as established. Whether or not $FeSiO_3$, as a crystalline compound with pyroxene structure, is thermodynamically stable is a question that cannot be answered with certainty at this time.

On the basis of the data for the ferromagnesian ortho- and metasilicates given above, the stability relation of enstatite and orthoferrosilite can be illustrated. Since the heat of solution of quartz in 20.1 per cent HF at 73°.7 C. has been measured by the authors ($\Delta H = -33{,}000 \pm 20$ cal/mol), the heats of the following reactions can be given:

1. $Mg_2SiO_4 + SiO_2 \rightarrow 2MgSiO_3$
2. $Fe_2SiO_4 + SiO_2 \rightarrow 2FeSiO_3$.

The entropies at 298°16 K. are known for Mg_2SiO_4 and $MgSiO_3$ (Kelley, 1943), for SiO_2 (Kelley, 1941*a*), and for Fe_2SiO_4 (Kelley, 1941*b*). For $FeSiO_3$ the entropy can be roughly predicted as $S^{\circ}_{FeSiO_3} =$ 23 cal/deg/mol. Accordingly, the heats and free energies of the two reactions mentioned above are, at 298°16 K.,

$$1.\ \Delta H = -2{,}620 \pm 280\ \text{cal.}\,,\quad \Delta F^{\circ} = -2{,}500 \pm 320\,;$$

$$2.\ \Delta H = -310 \pm 210\,,\quad \Delta F^{\circ} = -460\,.$$

Because of the great uncertainty in the entropy increment of reaction 2 and, on the other hand, of the small heat of that reaction, no reliance can be placed in the sign of the free energy of that reaction. The difference in the stability of $MgSiO_3$ and $FeSiO_3$ is, however, evident.

TABLE 3

HEATS OF SOLUTION OF $FeTiO_3$, $MgTiO_3$, Fe_2SiO_4, AND OF THE EIFEL OLIVINE IN AN ACID MIXTURE CONTAINING 10.05 PER CENT HF AND 10.05 PER CENT HCL AT 73°7 C.

Sample	ΔH (Cal/Mol)	Uncertainty (Cal/Mol)
Synthetic ilmenite. . . .	−46,650	60
Synthetic geikielite. . . .	−56,820	40
Synthetic fayalite.	−75,370	30
Forsterite, Dreis, Eifel, Germany.	−90,530	120

STABILITY RELATION OF ILMENITE AND GEIKIELITE

The abundance relation of ilmenite and geikielite very strongly favors ilmenite. Geikielite is extremely rare and has been found so far only in the gems of Ceylon.

To reveal the thermodynamical stability relation of ilmenite and geikielite in the presence of olivine or orthopyroxene, the heat-of-solution measurements listed in table 3 were undertaken. Because completion of the reaction of $MgTiO_3$ with pure HF proved to be too slow, an acid mixture containing 10.05 per cent HF and 10.05 per cent HCl was employed. The reported heat-of-solution values represent averages of the individual runs.

On the basis of the heat-of-solution values given in table 3 for synthetic fayalite and for Eifel olivine, the heat of solution of the forsterite-fayalite series in the acid mixture can be expressed by the following equation:

$$\Delta H = -75{,}370 - 165.2X_{mg}\,.$$

Extrapolation to pure Mg_2SiO_4 yields

$$\Delta H_{MgSiO_4} = -91{,}890\ \text{cal/mol}\,.$$

These heat-of-solution data yield, for the reaction $Fe_2SiO_4 + 2MgTiO_3 \rightarrow Mg_2SiO_4 + 2FeTiO_3$, the heat, $\Delta H = -3{,}820 \pm 250$ calories at 298°16 K. The entropies for Fe_2SiO_4 and Mg_2SiO_4, as given by Kelley (1941*b*, 1943), and for $FeTiO_3$ and $MgTiO_3$, as given by Shomate (1946), yield $\Delta F = -4{,}510$ calories as the free energy of the reaction at 298°16 K. This shows that the association $Mg_2SiO_4 + 2FeTiO_3$ is thermodynamically stable at room temperature. The negative free energy persists also at higher temperature ranges.

To approach the conditions prevailing in natural rocks, the following general reactions were considered.

$$1.\ n\,(Fe_{1.00-x}Mg_x)_2SiO_4 + MgTiO_3 \rightarrow n\,(Fe_{1.00-p}Mg_p)_2SiO_4 + FeTiO_3\,,$$

$$2.\ n\,(Fe_{1.00-x}Mg_x)\,SiO_3 + MgTiO_3 \rightarrow n\,(Fe_{1.00-p}Mg_p)\,SiO_3 + FeTiO_3\,.$$

In these reaction equations, n and x are independent variables, and p depends stoichiometrically on n and x.

Under the condition that the isomorphism of the forsterite-fayalite and the enstatite-orthoferrosilite series is thermally perfect, as indicated by the heat-of-solution data, the heats (ΔH) and the heat-capacity increments (ΔC_p) of the two reactions are calculable and are independent of the numerical values of n and x. The following high-temperature heat capacities were adopted: $MgTiO_3$ and $FeTiO_3$, given by Naylor and Cook (1946); Fe_2SiO_4 and $MgSiO_3$, given by Kelley (personal communication); Mg_2SiO_4 and $FeSiO_3$, estimated by means of the following approximations. $C_p(Mg_2SiO_4) = C_p(MgSiO_3) + C_p(MgO)$ and $C_p(FeSiO_3) = C_p(Fe_2SiO_4) - C_p(FeO)$. On the other hand, the entropy increments (ΔS) of reactions 1 and 2 depend largely on the values of n and x, because of the entropy of mixing of the corresponding isomorphic mixtures of the olivines and orthopyroxenes. The entropy of mixing was calculated according to the relationship, $\Delta S_{\text{mixing}} = -RN_1 \ln N_1 - RN_2 \ln N_2$, in which N_1 and N_2 are the mol fractions in the isomorphic minerals in question.

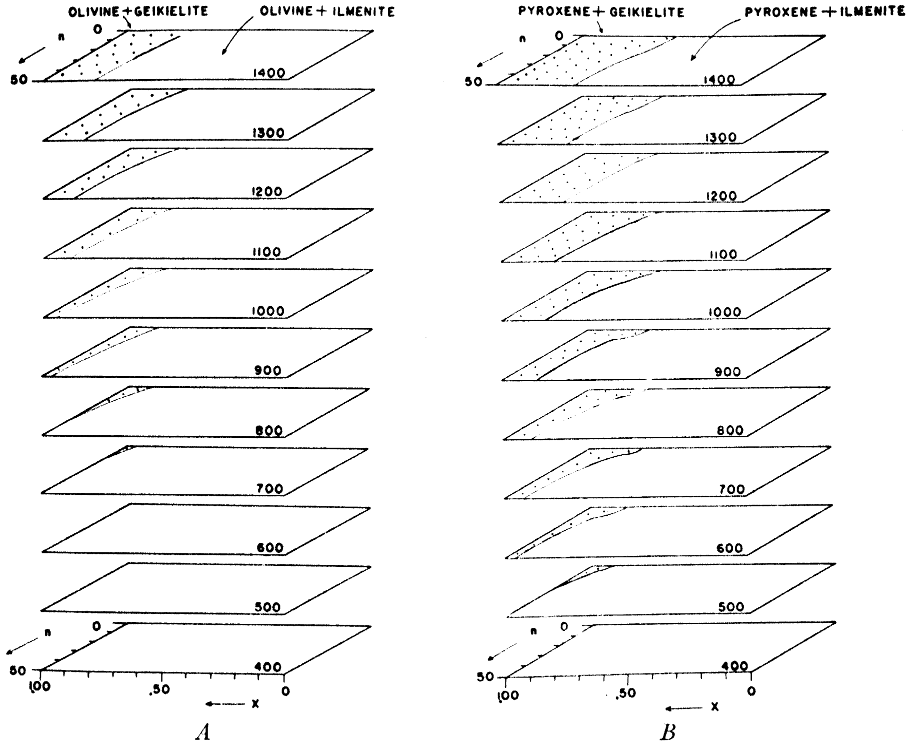

FIG. 2.—Stability relation of geikielite and ilmenite in the presence of olivine (A) or orthopyroxene (B) in a temperature range of 400°–1400° K.; x, mol fraction of magnesium in olivine (or pyroxene); n, molecular ratio of olivine (or pyroxene) to geikielite (or ilmenite) in the rock. The dotted area indicates the stability field of geikielite in the presence of olivine or pyroxene.

The free energies for the two general reactions 1 and 2 were computed from the data. The computations were made according to the relationship $d(\Delta F/T) = -(\Delta H/T^2)dT$. The thermodynamic nomenclature follows Lewis and Randall (1923). The results of the computations are shown diagrammatically in figure 2, *A* and *B*. The stability fields of geikielite in association with olivine or orthopyroxene, respectively, are indicated by the dotted areas on the individual sections corresponding to the different temperatures from 400° to 1,400° K.

Some of the thermal data employed in the above computations are approximations that may deviate more or less from the true values. Therefore, figure 2 can be applied only qualitatively or semiquantitatively. However, it shows that the association geikielite + olivine (pyroxene) represents a high-temperature paragenesis. In the presence of olivine or orthopyroxene, geikielite becomes stable only at higher temperatures and in rocks very rich in magnesium. This result indicates that geikielite may occur in those rocks more frequently than is known at present. Its proper identification in rocks is difficult and is possible only through separation and chemical tests.

Independently of the particular stability of geikielite, the present investigation shows that the entropy of mixing of an isomorphic mixture is a factor of geological importance and one which, in suitable instances, may greatly affect the stability relations of the minerals and mineral parageneses.

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ISOSTASY—EXTENDED[1]

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ABSTRACT

Basic principles are established for the equilibrium of a strong elastic lithosphere supported by a weak underlying magma. The lithosphere acts to spread out localized geological loads over a considerable area, and it is shown that isostatic equilibrium is achieved only when a load is reasonably uniform and distributed over a distance of some 330 km. This extended principle of isobaric equilibrium contains the isostatic principle as a special case. Basic mechanisms and methods for determining the deformation of the lithosphere under various kinds of loading are outlined. The geological significance of gravity anomalies is pointed out, and methods of forecasting their magnitude and distribution in terms of the deformation of the lithosphere are given.

The extended theory of isostatic equilibrium that the author has called "isobaric equilibrium" finds application in all quantitative discussions of the deformation of the lithosphere.

INTRODUCTION

In a series of papers appearing for the most part in the *Journal of the Franklin Institute* and in *Geophysics*, the author has presented an extension of the principles of isostatic equilibrium (1937, 1943*a*, *b*, 1944, 1945, 1947). This extension deals particularly with the part played by a strong elastic lithosphere and is especially significant in areas exhibiting crustal deformation. Because these early papers were mathematical and not entirely understood by many readers, it is the object of the present paper to review the concepts and present them in a simpler and more usable form. For formal justification of some of the conclusions outlined herein, the reader may therefore find it desirable to refer to the original publications.

Modern ideas concerning the equilibrium of the lithosphere were introduced in 1854, when J. H. Pratt (1855) prepared an extensive report on the figure of the earth and gave an explanation for the apparent deficiency in the gravitational attraction of mountain masses as measured in closely adjacent areas. Shortly thereafter, George B. Airy (1855), in discussing the discrepancy brought out by Pratt's work, wrote: "It appears to me that the state of the earth's crust lying upon the lava may be compared with perfect correctness to the state of a raft of timber floating upon water; in which, if we remark one log whose upper surface floats much higher than the upper surfaces of the others, we are certain that its lower surface lies deeper in the water than the lower surfaces of the others."

The various ideas of Airy and Pratt have been examined in the light of many observations, and these have established the principle of isostasy, at least for large, relatively undeformed crustal areas. William Bowie (1931), one of the active workers in this field, defines isostatic equilibrium as

> the condition of rest which the outer materials of the earth tend to acquire. The equilibrium is between the crust and the subcrustal material and *not within the crust itself* [italics by the author]. It is based on the idea that the mass of each unit section of the earth's crust exerts the same pressure on the subcrustal material. According to the principle of isostasy, if the earth's crust were cut into prisms of the same cross-section by imaginary vertical planes, the prisms would have the same mass if the isostatic condition were perfect.

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This formulation of the isostatic principle, which specifically ignores the strength of the lithosphere, has guided geological interpretation for thirty-five years. The principle, as stated, is an incomplete formulation of crustal equilibrium, and the whole problem needs reconsideration.

As a general rule, the differences between the calculated and the observed values of the acceleration due to gravity, or the isostatic anomalies, in undeformed and level country and over most of the oceans are relatively low, being equivalent, on the average, to a layer of rock about 400 feet thick. However, in recent years, as methods for correcting for elevation, density, and topography have become more accurate and detailed data from the field have become available, it has become increasingly clear that many regions of the earth cannot possibly be in isostatic adjustment. Large gravity anomalies have been interpreted as indicating gross departures from normal density in the vicinity of the observing stations and frequently have led to misleading statements suggesting that the anomalies imply an unstable geological structure, even though clear-cut observational evidence is available to show that the region is entirely stable and that no diastrophic activity is in progress. Such discrepancies in the interpretation of geological structure in regions of gravity anomaly became so glaring, particularly in relation to theories of mountain-building, that the author was led to review and extend isostatic theory (1937–1947) with the object of bringing the common gravity irregularities into the framework of natural physical law. This rationalization was accomplished by including an aspect of crustal stability specifically ignored in Bowie's foregoing definition of isostasy by quantitatively formulating the effect of finite strength and stresses *within the crust itself*. These neglected factors are found to be of fundamental importance.

IDEAL ISOSTASY

Consider a newborn earth made up of rock material at such a temperature that its strength in shear is negligible. After circulation and mixing have acted for a few million years, the underlying layers of such an earth will approach the hydrostatic state characterized by (*a*) a steadily increasing temperature and density toward the center; (*b*) the coincidence of surfaces of equal density, equal pressure, and equal gravitational potential; and (*c*) a surface layer of cool and strong rock, whose density is notably less than the underlying hotter rocks.

Suppose that great prisms of cool rock are ultimately formed at the surface by selective crystallization or some other means and that these aggregates are systematically less dense than the underlying weak magma. Obviously, they will be supported according to Archimedes' hydrostatic principle and will float in such a way that the weight of each prism will be exactly equal to the weight of the displaced underlying magma.[2] Consider a number of prisms supported in the above manner, all of different but comparable density. Assume that each prism is self-supported and that *no friction whatever exists between the lateral faces.* Then, if a reference prism is selected and if the acceleration due to gravity is measured on it and if suitable corrections are made for the distance of the observing station above some reference level and for the figure of the earth, it will be found that the measured values on all prisms will not appreciably differ. This is a consequence of the laws of gravitational attraction and potential theory, which show that to a sufficient approximation the vertical acceleration due to a horizontal plate of considerable extent is given by the familiar relation

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[2] The term "magma" is here used to denote any type of underlying rock having inappreciable strength in shear.

$$\Delta g = 2\pi\gamma M\ , \qquad (1)$$

where Δg is the change in gravitational acceleration due to the plate, M is the mass per unit area of the plate, and γ is the gravitational constant. If the plate is not extensive, geometrical corrections must be applied, but equation (1) will represent the principal term. Since, by Archimedes' principle, the mass per unit area of the floating block is, on the average, exactly equal to the mass per unit area of the displaced liquid and since by equation (1) the gravitational defect depends on this quantity, it is clear that different thicknesses or densities of a floating block will not produce important changes in the measured value of the acceleration due to gravity.

Now, if a layer of sediments be piled on top of one of the prisms by transfer from some other place, the prism will sink until it displaces a weight equal to that transferred to its surface; and, if sediments are washed off the prism, it will rise in a strictly analogous manner. Thus, as long as the prisms are perfectly free to move vertically in relation to one another, the basic ideas expressed by Airy apply. Because such a system of freely floating blocks will show a constant gravitational acceleration irrespective of the sedimentary load placed thereon, we may describe the state as one of ideal isostasy. The principal term of the gravity anomaly for such configurations is zero. Although the commonly used statement that the mass per unit area above a given equipotential surface is constant is sufficiently accurate for most cases of *ideal isostasy*, yet occasionally the exact statement must be employed (Gunn, 1936), namely, that *the mass per unit area multiplied by mean gravity in the prisms, is a* CONSTANT.

DEPARTURES FROM IDEAL ISOSTASY

In order to understand the basic mechanisms, consider a simple experiment with the frictionless prisms mentioned above. It has been noted that adding or subtracting loads of sediments to the prisms will not influence the gravitational acceleration as measured at a standard level on these prisms. Suppose, now, that, instead of a load of sediments, a large force of unspecified origin acts downward, which will displace from underneath the prism a certain mass of magma whose weight by Archimedes' principle is exactly equal to the applied downward force. Because no material is transferred to the block and because mass has been forced out from under it, it is clear that a deficiency of mass exists in this prism in relation to the initial condition of ideal isostasy, and therefore a negative gravity anomaly will be measured. The magnitude of this gravity anomaly is given by equation (1), where M is the mass per unit area of the magma *displaced* by the downward-applied force. In a similar way a vertically upward force might be applied to an adjacent prism, resulting in an inflow of excess mass below the prism in such a way that a positive gravity anomaly would be measured on this prism. Although we cannot conceive of such disconnected geological forces that will depress or lift frictionless prisms of the earth's crust, it is entirely possible for vertical forces to be applied to each prism arising from the frictional contact with and motion of adjacent prisms. Such forces could be very important. It is worth while, therefore, to go one step further and assume that, instead of the prisms being frictionless, they are actually in elastic and cohesive contact with their neighbors, forming a great aggregate of prisms and constituting an elastic sheet having great internal strength. Frictional and cohesive forces between the vertical faces now prevent the individual adjustment of the prisms. For example, select a central prism and assume that sediments are piled on it alone. If this prism stood alone and was supported only by the weak asthenosphere, it would sink in the substratum and displace a weight of material exactly equal to that added. However, because of the assumed forces of friction or cohesion acting on the vertical faces of the prism, this cannot take place. The applied load attempts to force the prism down, but now, because it is mechanically associated with the neighboring prisms, the latter are forced down while the original prism is supported by shearing forces to exactly the same extent. Thus, under the selected prism not enough mass is forced out laterally to support it from the bottom, and part of the support is supplied by vertical shearing forces appearing on the vertical faces of the prism. Conversely, as a result of such forces, considerable material is forced out from under the adjacent prisms, even though no load of sediments is applied to them. Of course, considering the aggregate of prisms, just enough material is displaced to carry the superposed load of sediments on the original prism, and a specialized condition of regional isostasy is achieved.

The deficiency or excess of mass displaced from under any prism can generally be calculated in terms of the applied force or the vertical stress. For simplicity, consider a prism having frictionless walls and of unit cross-section supported ideally by the substratum. Suppose this prism is pushed down by some arbitrary vertical force acting on unit area or by shearing forces arising from the finite strength of the lithosphere. Let this force per unit area be the vertical resultant supporting stress, S_v. Such a stress, acting on a freely moving prism, will displace substratum material until a restoring stress, equal to the weight of the displaced liquid, restores equilibrium, or

$$S_v = M'g \,, \qquad (2)$$

where M' is the mass per unit area of the *displaced* material and g is the acceleration due to gravity. It is clear, therefore, that a downward or positive resultant stress from any source whatever results in a deficiency of mass under the prism, while an upward stress results in an excess of mass. If this change in mass is assumed to be in the form of a large flat sheet, the principal term of the accompanying gravitational anomaly, Δg, increases with the applied stress; or, specifically by equations (1) and (2), this is given by

$$g\Delta g = 2\pi\gamma S_v \,. \qquad (3)$$

The frictional or cohesive forces acting on the vertical faces of the prisms necessarily produce downward (or upward) forces in adjacent prisms, and these prisms are just as much in equilibrium as if they were supported ideally from the bottom. It follows from the above considerations that, if the resultant vertical stress distribution can be determined for any arbitrary load on the lithosphere either by measurement or by calculation, then it is possible to evaluate the gravity anomalies associated therewith. Conversely, a knowledge of the space distribution of gravity anomaly and mass permits an estimate of the vertical superposed stresses and the resulting deformations of the lithosphere.

Because the integrated upward and downward stresses over a section of the lithosphere more than 400 km. long must add up to zero, it follows that the space integral of the principal term of the gravity anomaly over the same path must also approach zero.

THE LITHOSPHERE AND ITS SUPPORTING MAGMA

The lithosphere may be defined as the cool, strong, rocky layer of the earth that is supported by the denser, weaker, and relatively hotter underlying layers. The upper few kilometers of the lithosphere are traversed by many shallow faults and fractures. As a unit, these upper layers are relatively weak, but at a depth of perhaps 5 km. and more, where the compressive stress due to the superincumbent rocks exceeds the crushing strength, local fractures of the type regularly occurring at the surface itself become more or less self-healing, and this part of the lithosphere may be considered to be equivalent to a strong, uniform elastic sheet. By "strong," the author means that the crustal material will support shearing stresses of 1 or 2 $\times 10^9$ dynes/cm^2 for periods of time comparable to the age of mountains.

It is a convenient mathematical fiction to assume that the strong lithosphere extends down to a definite level and that below this level the supporting magma is fluid in nature and has zero strength in shear. Actually, the transition from the strong lithosphere to the weak magma is continuous. The existence of high mountains and ocean deeps shows that rocks just below the surface are strong and will support large shearing stresses for long periods of time. But as the depth increases, the temperature rises steadily; and, even though it is possible that the shear strength for short impulses is maintained, yet at sufficiently elevated temperatures progressive plastic deformation will almost certainly be produced by stresses applied for a sufficiently long time. Too little is known of the mechanical properties of great sheets of rock, but there is accumulating evidence to show that the apparent shearing strength of geological materials decreases as the time of application of the stresses increases and that both the strength and the time interval necessary to induce appreciable creep decrease notably as the temperature increases. Thus the steady increase in temperature as one goes deeper and deeper into the crust implies a continuous change in the apparent strength of the rocks. An examination of the continuous transition to the weak magma in comparison with a discontinuous one shows that, except for some uncertainty in the position of the neutral axis, an elastic outer surface of the earth so constituted behaves mechanically in a manner remarkably like a strong, uniform, elastic lithosphere of perhaps 50-km. thickness, supported by and superposed upon a weak magma. A constant that measures the ease of deformation of the lithosphere (characteristic constant = *b*) is found to be by far the most important quantity describing its mechanical characteristics, and this constant is related to an equivalent thickness that becomes the true thickness only when the transition from the lithosphere to the asthenosphere is assumed to be discontinuous. It is sufficiently accurate, therefore, to assume that the lithosphere is a strong, elastic sheet of sufficient uniformity that it obeys well-established laws of elasticity and that it is everywhere supported on a weak underlying magma that, given adequate time, will adjust itself to hydrostatic equilibrium. This time necessary for adjustment is short in most cases of geological interest and probably approximates 25,000 years.

ISOBARY—A BASIC EQUILIBRIUM PRINCIPLE EMBRACING ISOSTASY AS A SPECIAL CASE

Because it is assumed that the lithosphere is everywhere supported on a substratum having negligible strength in shear, it is possible to establish the criterion for vertical equilibrium in any selected prism. In the asthenosphere below the lowest section of the lithosphere, some surface of equal pressure or equal gravitational potential is adopted as a reference. If an applied vertical load on the lithosphere should increase the pressure at this selected level, hydrostatic laws show that transient horizontal pressure gradients will be set up, and these will force the weak rock outward more or less horizontally until the constancy of pressure over the selected equipotential surface is re-established. It is a property of the hydrostatic state that, when constancy of pressure is re-established, the system is in equilibrium. Now the pressure at this selected level is made up of three components, namely: (*a*) the weight per unit area of the column of weak rock extending from the reference level up to the bottom of the lithosphere; (*b*) the weight per unit area of the column that extends from the bottom of the lithosphere to the actual surface; and (*c*) the vertical stress, S_v, imposed on the column by the finite strength of the lithosphere. The total pressure arising from components *a* and *b* may be conveniently written $M\bar{g}$, where M is the mass per unit area lying above the reference equipotential surface and $\bar{g}$ is the average gravitational acceleration in any selected prism.

The criterion for isobaric equilibrium, therefore, is

$$P = M\bar{g} + S_v = \text{Constant}\,, \qquad (4)$$

where P is the pressure at the selected equipotential or isobaric level and S_v is the superposed vertical stress. This relation differs from the analogous relation for isostasy by the inclusion of the vertical superposed stress, S_v. The relation is generally applicable, even though each prism may be elastically attached to its neighbors. The quantity S_v is conveniently taken to be positive when it acts in the direction of $\bar{g}$ or downward. This exact result may be stated in words as follows: For sections of the lithosphere to be in stable or isobaric equilibrium, it is necessary and sufficient, in prisms reaching from a selected surface of equal pressure in the asthenosphere to the actual surface, that the *vertical force per unit area* arising from the finite strength of the lithosphere, added to the product of the *mass per unit area* and the average acceleration due to *gravity, shall be a constant.*

When the lithosphere is adjusted to this condition, one may conveniently say it is in "isobaric equilibrium," since the word "isobaric" implies equal pressure. This extended formulation of isostatic equilibrium has been called "isobaric equilibrium" following a suggestion of C. E. Dutton, who originally discarded this more descriptive term in favor of "isostasy." Isobary, therefore, corresponds to a state of vertical mechanical equilibrium, even though it permits real anomalies of gravity to exist. In level regions where vertical stresses are negligible, isobary obviously reduces to the more restricted equilibrium of isostasy. It is convenient to pick some standard column as a reference, which should preferably be near the center of a large undeformed region. At such a place one may be assured that the superposed vertical stresses usually due to deformation are small and the regional isostatic adjustment approaches the ideal, quite irrespective of the finite strength of the lithosphere. Now, if gravitational measurements at other places are compared with those of such a reference point, making appropriate corrections for its position in relation to the geoid and the figure of the earth, it is clear that gravity anomalies will be an indication of vertical stress. Equation (3) may be combined with equation (4), and the principle of isobaric equilibrium may be written in the approximate but convenient form,

$$P = M\bar{g} - \frac{g\Delta g}{2\pi\gamma} = \text{Constant} \,, \qquad (5)$$

where Δg is the principal term of the gravity anomaly. The exact anomaly may be calculated if the geometrical distribution of stress, mass, and topography is available in any given case. The principle of isobaric equilibrium as here expressed, together with observed numerical values for the gravity anomalies, provides a valuable tool for the study and determination of lithospheric deformations, particularly in special geological structures that are definite exceptions to the principle of isostasy as usually defined.

DEFORMATION OF A HYDROSTATICALLY SUPPORTED ELASTIC LITHOSPHERE

In order to achieve a basic familiarity and understanding of the types of equilibrium that exist when a mountain is supported by isostatic forces and when it is supported by isobaric forces, it may be illuminating to introduce an analogy at this point. Suppose a small boy walks out on a deep pond covered with thin ice. If the ice is too thin, he breaks through, and, according to Archimedes' principle, he is then supported in the water only to the extent of the weight of water that he displaces by his body. Quite obviously, the support is in accordance with hydrostatic principles, and, assuming he has a normal density, complete support is not achieved until he is almost immersed. Now, if the thickness of the ice increases, a thickness is finally reached which will support the boy. This thickness corresponds to the condition that the weight of the boy is distributed over a big enough area that the actual depression of the ice below its normal level is sufficient to displace a weight of water exactly equal to the weight of the boy. It might be said that, when the ice is strong enough, its downward depression by the boy produces a very shallow circular boat which displaces enough water to support him effectively. This downward deflection of relatively thin ice is more or less apparent to every skater, and it is clear that the strength and elasticity of the ice plays an important part in the support of the skater. Now, as the thickness of the ice increases, its flexural rigidity also increases, and the support for the weight of the boy is spread out over a greater and greater area. Simultaneously, the magnitude of the systematic downward deformation of the ice due to the weight of the boy is so reduced that the weight of displaced water is maintained exactly equal to the weight of the boy. It should be clear, further, that the weight of the skater is never passed back to the supporting banks of the pond but is always passed downward into the supporting and underlying fluid. Whether the boy is on the ice or in the water is immaterial in relation to the amount of displaced underlying fluid, but it is clear that the distribution of the displacement in the two cases is radically different. It should be evident that the strength of the ice plays an important role in modifying the geometry of any given system of loads. The analogy is such that the boy floating in the water corresponds to the floating equilibrium type of support envisaged by isostasy, whereas the distribution of displacement of the supporting water and the stress in the ice sheet correspond to isobaric equilibrium. In the language of isostasy, the "compensation" for the boy's head, which is above water, is to be found immediately below in the deficiency of density of the boy's body. On the other hand, when the boy is on top of the ice sheet, the compensation for his weight is spread out over a considerable area by the strength of the ice and is to be found in the water displaced and pushed aside by the downwardly deformed ice sheet. Attention is drawn to the fact that the ice even some distance from the boy is bent downward well below its equilibrium position, and the mass of water in a typical unit area is accordingly less than normal. Thus, if a finite strength is assigned to the lithosphere, most of the compensation for a given load is not under the load but to the sides. This is in sharp contrast to the basic assumptions of isostasy.

Analysis shows that, with knowledge of the thickness and elastic modulus for the ice and the density of the underlying fluid, it is possible to calculate, using straightforward engineering principles, the magnitude of the downward deformation of the ice for all points immediately under the boy and in his vicinity. It is found that the volume of the depression caused by the boy, multiplied by the density of the water, is exactly equal to the mass of the boy. This is true for any thickness of ice. The dimensions of the depression and the effective radius over which the boy's weight is distributed can also be determined.

Although the above basic ideas underlying isobary are very simple, it is evident, in order to calculate the vertical supporting stresses arising from a superposed load on the actual lithosphere, that some rather detailed calculations must be carried out. The deflection produced in the lithosphere by a single large vertical force may be computed with relative ease and is strictly analogous to the case of the small boy standing on a sheet of ice. There are practical mathematical difficulties involved in computing the downward deflection of the lithosphere for loads distributed over a considerable area. An investigation has shown, however, that if the distributed load is linear in character and extends for some hundreds of kilometers in such a way that the load distribution can be well represented by a single section taken at right angles to the linear trend, then the downward deformation for a number of distributed loadings may be easily worked out. Now, because many very important earth features like deeps, mountain chains, and continental borders are essentially linear in character and extend for a thousand kilometers or so, this mathematical simplification is adequate for most geological interpretations and enables one to work out in certain areas the exact deformed figure of the lithosphere.

In order to determine the figure for any given loading, it is convenient to select a typical section of unit width at right angles to the linear trend and consider this slice of the lithosphere as a strong elastic beam resting on, and ultimately supported by, the weak underlying magma. It is assumed that Young's modulus of elasticity and the sectional moment of inertia per unit width is substantially constant throughout the explored region. The reader is referred to the original papers for the details of such calculations. It may be stated, however, that, once the load applied to the original undeformed lithosphere is specified or known, it is practicable by the solution of the basic equations to determine at once (*a*) the downward or upward displacement of the lithosphere, relative to its position prior to the application of the load, for all points along the section; (*b*) the distribution of the internal bending moment and the vertical shear; (*c*) the vertical restoring force per unit area, S_v, due to the deformation of the lithosphere, which is, in fact, proportional to the space gradient of the vertical shear; (*d*) the fiber stress at every point in the equivalent beam, and (*e*) the principal term of the gravity anomaly resulting from the internal supporting stresses. It is evident that all the foregoing items are of considerable importance in structural studies of the lithosphere; and because these may be determined, once the exact figure of the lithosphere is known, practical methods for its evaluation seem important.

The deformation of the lithosphere considered as an elastically supported beam was first carried out by M. Smoluchowski (1909), who particularly considered solutions for the deformation on the assumption that the crust was composed of successive layers of rock and that its flexural rigidity was very much lower than is now believed to be the case. The roots of the basic equations for this situation lead to periodic solutions that this author thinks could have been descriptive of the lithosphere as a whole only in pre-Cambrian time, when the crust was very thin and weak. The roots of the basic equations employed in the author's investigations are those corresponding to a homogeneous and elastically strong lithosphere, and these lead to logarithmically attenuated waves, which well describe actual geological formations. It can be shown (Gunn, 1937–1947) that the solution to the basic equations having important geological applications leads to vertical deformations, y, of the form

$$y = P\epsilon^{-ax} \sin bx + Q\epsilon^{-ax} \cos bx \ , \quad (6)$$

where, in most practical cases, a is indistinguishable from the characteristic constant b, while P and Q are amplitudes depending on the imposed boundary conditions and the mechanical characteristics of the lithosphere. The vertical deformation y given by equation (6) may be represented for any assumed vertical force by a simple curve, as shown in figure 1. In figure 1, *A*, the downward deformation of the lithosphere due to a downward force is shown for an unbroken section and in figure 1, *B*, for the same force applied to one end of a broken section. The abscissae are expressed in terms of an angle measured in radians, which is the product of the characteristic constant b and the distance x along the beam from the applied force. In the figure actual distances are also plotted for the earth.

It is important to note from this figure that in the actual earth the load applied at a specific point appreciably deforms the lithosphere for a distance of some 250 km., quite irrespective of how large that force may be. As the force increases, the downward deformation increases proportionately, and the resulting deformation is increased at every position along the beam by the same factor. The distance from the force to the point on the lithosphere where the downward deformation is zero is independent of the force and is measured only by the flexural rigidity of the beam and the density of the underly-

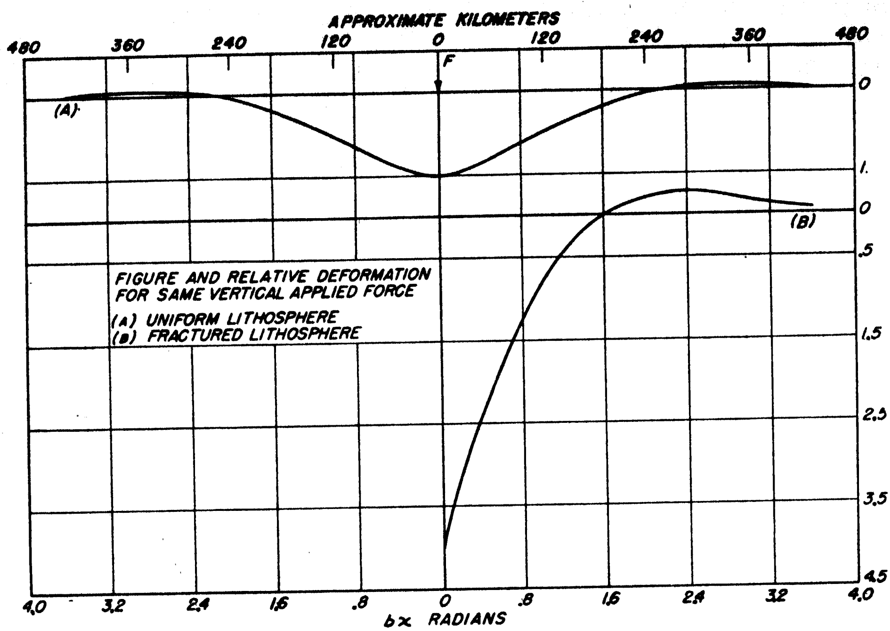

FIG. 1.—Characteristic deformation produced by a single vertical force

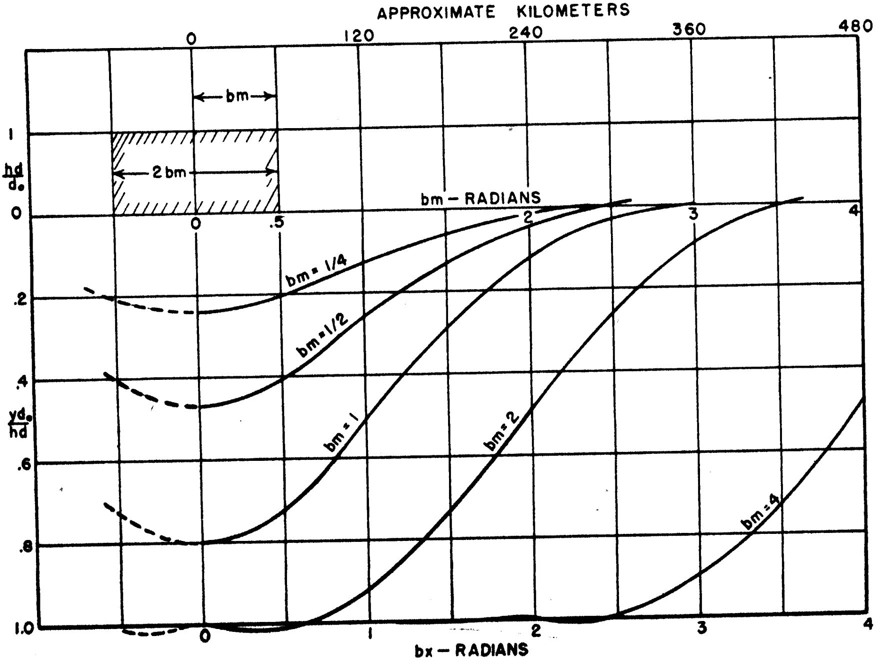

FIG. 2.—Deformation produced by long block loads of various widths

ing magma. These, in turn, determine the value of the characteristic constant b that measures the ease of deformation of the lithosphere.

The actual deformation produced by any type of load can be evaluated by replacing the force, F, in figure 1, A, by an elementary load and summing up all the downward displacements caused by each element. Thus, if a selected load of any configuration is divided up into a series of narrowly defined loads and the downward displacement of each is calculated, the total downward deformation is evidently the sum of all the deformations, due allowance always being made for the exact position of each element. Certain geometrical forms are easily calculated, and a number of these have been presented in graphical form for reference and use (Gunn, 1937–1947). Figure 2 shows the deformation of the lithosphere that results from a long block load of width 2 bm, height h, and density d, as calculated by the above method. The figure brings out the fact that, as the width of the load increases, the block forces the lithosphere deeper and deeper into the underlying magma until at a width of the block equal to something like 4 bm the block is essentially submerged to a level that approaches the isostatic condition. The degree of support provided by the lithosphere and its influence on isostasy and gravity anomalies as expressed in terms of the size of the block will be discussed in a later section. In order that the plotted data may be of maximum usefulness, in figure 2 and in the following figures distances along the beam are measured in radians (bx), so that the curves may be generally applied to any section of the lithosphere corresponding to different values of b, its characteristic constant. The normal value of b in the earth's crust is such that the distance for one radian as plotted in this series of curves closely approximates 120 km., so that an approximate distance scale may be simultaneously plotted on each curve. For example, in figure 2, the semi-width $bm = \frac{1}{4}$ corresponds to a width equal to one-fourth of 120 km., or 30 km.; and we see that such a block, of total width 60 km., is deformed downward into the underlying magma by a distance equal to 0.24 of the equivalent height of the block. Thus, by knowing the equivalent height of a geological load having the form of a long block and the appropriate numerical factor as determined from figure 2, the exact downward deformation in kilometers may be immediately estimated.

The author's papers give a large number of figures showing the downward deformation of the crust produced by various loads. It is worth while to note that almost any load can be simulated with sufficient accuracy by the addition of these standard deformations. These deformations superpose, and it follows from the mathematical form of the equations that the sum of the lithospheric deformations due to each load is the same as the deformation for the sum of the loads.

THE SIGNIFICANCE AND MAGNITUDE OF THE CHARACTERISTIC CONSTANT b

The flexural rigidity of the lithosphere and the density of the underlying weak magma, together with the actual value of acceleration due to gravity, determine the magnitude of the characteristic constant of the lithosphere in such a way that, when its value b is multiplied by the distance x along the effective beam, the resulting product is the angle bx expressed in radians. The quantity b, therefore, is the reciprocal of the relaxation distance of the lithosphere, and this may be defined as the approximate distance required for gross deformations to be appreciably attenuated. This characteristic distance ($1/b$) approximates 120 km., and b is $(8.4 \pm 2)\ 10^{-8}\ \text{cm}^{-1}$.

The numerical value of b may be calculated in a straightforward manner in terms of the roots of the basic equations which define b if the lithosphere is reasonably homogeneous. More specifically, b is equal to the fourth root of the product of the density of the underlying magma (diminished by the density of any sediments or water that washes into the depression) multiplied by the acceleration due to gravity and divided by four times the flexural rigidity of the lithosphere, or

$$b = \left[\frac{(d_0 - d)\,g}{4\ (\text{flexural rigidity})}\right]^{1/4} = \left[\frac{3\,(d_0 - d)\,g}{ET^3}\right]^{1/4}, \qquad (7)$$

where E is Young's modulus of elasticity, T is the effective thickness of the strong lithosphere, d_0 the density of the underlying weak magma, d the density of water or sediments that may *completely fill* any downward deformation, and g the acceleration due to gravity. Thus b is large for a very weak and yielding crust or one in which plastic flow occurs, but it approaches zero for a crust that is perfectly rigid. The quantity b, therefore, measures the ease of deformation of the crust by a given load. The correct values for the above quantities are not exactly known. For example, the effective value of Young's modulus in the earth's crust according to competent investigators has been taken as low as 2×10^{11} dynes/cm^2 to as high as 2×10^{12} dynes/cm^2. The thickness of the crust has similarly been given as from 30 to 90 km. Taking the outside limits above and assuming $d_0 = 3.1$ gm/cm^3, $d = 0$, and $g = 980$ cm/sec^2, it is easy to show from equation (7) that the smallest probable value for b is $5.45 \times 10^{-8}\ \text{cm}^{-1}$ and the highest is $20.2 \times 10^{-8}\ \text{cm}^{-1}$. For the present, one may take E as close to 1.2×10^{12} dynes/cm^2 and $T = 50$ km. Substitution of these in equation (8) leads to $b = 8.4 \times 10^{-8}\ \text{cm}^{-1}$.

Because the above direct calculation of the characteristic constant b is open to uncertainty and because its value is of first-rate importance in all discussions of the stress and deformation of the lithosphere, it is of importance to examine various known geological structures to see how observed b changes under widely varying terrestrial conditions. In order to secure reliable values, it is necessary to select for analysis simple geological structures in which the acting forces are both known in position and localized in nature. As an example, consider the downward deformation produced by a localized load, as shown in figure 1. The sediments initially deposited at the single mouth of a great river might well represent such a concentrated load, and, instead of these sinking at the point of deposition, thus establishing isostatic equilibrium, the strength of the lithosphere supports the sediments locally and passes the load to adjacent regions. Now, as more sediments are deposited by the river, they are spread out in a great delta, producing a downward deformation like that of figure 1, *A*, and having a total width of $bx = 3\pi/2$. This basin of widely deformed ocean bottom encourages the deposition of sediments at a considerable distance from the original mouth, and a great, symmetrical delta is built up. The mechanical behavior of the crust suggests that sedimentation will proceed laterally until the downward deflection is filled, at least thinly, to the edge of the original depression. When such a delta is well developed, as in the case of the Nile and Niger rivers (Barrell, 1914), the extent and figure of the delta describes the extent and approximate figure of the underlying downwardly deformed lithosphere. Barrell's data show that the total width for both the above-mentioned deltas is about 600 km., from which it is easy to calculate that b is equal to 7.85×10^{-8} cm^{-1}, or the relaxation distance $1/b = 127$ km.

As is shown in the basic papers listed, it is possible to calculate the principal term of the gravity anomaly when the deformations of the lithosphere are known. The detailed characteristics of a highly compressed segment of the lithosphere that has failed in shear along some zone of weakness have been worked out, and the expected distribution of gravity anomaly therein has been calculated. These calculations (Gunn, 1937–1947) show that the overriding and underthrusting of the sheared ends produce positive anomalies in the overriding sector and negative anomalies in the underthrusting sector that extend for a distance away from the point of failure equal to an angular distance, $bx = \pi/2$. Beyond this negative zone in the underthrusting section is a positive area extending for a distance of $bx = \pi$. By the use of observed data obtained by Vening Meinesz in the Javanese area, b may be immediately calculated, and it is found that b for this area approximates 8.6×10^{-8} cm. Gravity data from a number of points on the earth, when they are summarized and allowance is made for geometrical variation and uncertainty, lead to a value for $b = (8.4 \pm 2) \times 10^{-8}$ cm^{-1}. Although the suggested values of b do not appear to change greatly over the surface of the earth, this matter deserves special investigation and may shed considerable new light on the mechanical characteristics of the crust.

VERTICAL SUPPORTING STRESS PROVIDED BY THE STRONG LITHOSPHERE—ISOBARIC STRESS FUNCTION

Because the underlying weak magma always approaches the hydrostatic state, the equilibrium of the lithosphere under an applied load is consummated when the pressure along a given gravitational equipotential surface is constant. Thus, on comparing the total pressure in a typical undeformed area with that where the lithosphere is notably distorted, one needs only to equate the expressions for the pressure at the equipotential surface in the two areas. The pressure in general is made up of two components, (*a*) the weight of the superincumbent rocks, including both those of the lithosphere and the load (where it exists), and (*b*) the vertical stress supplied by the strong and elastic lithosphere. If the lithosphere is bent downward by a given load, the pressure under the load at the originally selected reference level is exactly equal to the pressure some distance away in an undeformed area. It may be shown that, under this condition,

$$-S_v = h\,d\,g - y\,d_0 g = d_0 g\,(H - y)\,, \quad (8)$$

where H is the "equivalent height" at any position, y is the downward deformation immediately below, and S_v the vertical supporting stress contributed by the lithosphere. The equivalent height, H, is the height that any given load would have if it had the same density as the underlying magma. It is defined by

$$H = \frac{dh}{d_0}\,, \quad (9)$$

where d and h are the density and actual height of the load, respectively, and d_0 is the density of the underlying magma. The relation of equation (8) states, simply, that the downward force per unit area due to the applied load, hdg (or, better, Hd_0g), under conditions of equilibrium is exactly equal to the upward force per unit area due to the downwardly displaced magma, yd_0g, added to the upward force per unit area ($-S_v$) due to the strength of the lithosphere. Since S_v is taken positive in the direction of g (the acceleration due to gravity), S_v directly under a load is usually negative. The relation expressed by equation (8) is invaluable in all discussions of stresses in the lithosphere and gravity anomalies and is called the "isobaric stress function." By methods beyond the scope of this review, it is possible to express the vertical stress, S_v, in terms of a high-order curvature of the lithosphere rather than in terms of the load and displacement. It was shown in the earlier sections leading to equation (3) how the principal term of the gravity anomaly can be expressed in terms of the vertical stress so that one may finally write the multiple basic relations derived in the series of papers here outlined (Gunn, 1937–1947) as

$$H - y = \frac{EI}{d_0 g}\frac{d^4 y}{dx^4} = \frac{\Delta g}{2\pi\gamma d_0} = -\frac{S_v}{d_0 g}. \quad (10)$$

In these relations, one must not confuse the vertical stress, S_v, with the fiber stresses within the lithosphere that determine the internal elastic equilibrium but should note, rather, that $-S_v$ is the *supporting force per unit area* contributed by the strength of the lithosphere itself.

The above principles may be applied to give specific answers to perplexing problems concerning isostasy and the occurrence of gravity anomalies. For example, it has heretofore been impossible to state just how big a block load carried by the crust must be, in order for it to approach isostatic equilibrium. It should be clear from figure 2 that the downward deformation by a long block load expressed as a fraction of its equivalent thickness increases almost linearly with increasing width of the block up to about $2\ bm = 1$ (or of total width about 120 km.), while it is essentially submerged for widths greater than $2\ bm = 4$ (or of a total width of 480 km.). Thus block loads of limited extent are carried almost completely by the strong lithosphere, and the ultimate support is passed downward to the weak underlying magma over a large area, whereas extensive loads, because of their considerable downward deformation, are largely carried by the hydrostatic forces provided by displaced underlying magma. By the use of curves like that plotted in figure 2, it is possible to determine the fraction of the load carried by the strong lithosphere itself as well as that carried by the immediately underlying weak magma. The ratio of the supporting forces contributed by the weak magma immediately below to that provided by the strong lithosphere is important because it measures the approach of any given load to ideal isostatic adjustment. Thus, if the weak substratum carries all the load, the ratio is unity and ideal isostasy prevails, whereas if the load is carried entirely by the lithosphere and distributed to the substratum over a large area, the ratio is small. A plot of the fraction of the total load carried by the asthenosphere immediately below is given in figure 3 as a function of the semidimension of the block. It follows from this curve that, for a long block load to be even in approximate (96 per cent) isostatic equilibrium, its width must be not less than 330 km.—a value notably larger than the values frequently suggested in the literature of isostasy.

FIBER STRESS AND MECHANICAL FAILURE WITHIN THE LITHOSPHERE

The foregoing outline has presented curves showing how the downward deformation of the lithosphere under various types of loading can be determined, and one may, therefore, state the approximate position of the neutral axis relative to the original surface of the earth at every point. Thus, by using the standard methods of engineering mechanics, the fiber stress at every point inside the equivalent beam can be formally calculated. Actually, however, the graphical methods suggested above are not particularly adapted to this type of calculation, and analytical solutions are much more accurate. Because this type of analysis is beyond the scope of this simplified review, it is only remarked that, since the fiber stress within the lithosphere can be calculated for every point as a function of its distortion, then, with the knowledge of the strength of rocks that constitute the lithosphere, one may immediately estimate where it will break when subject to sufficient deformation. Thus, for example, in a study of the Cape Henry area (Gunn, 1944) it was possible to calculate with a fair degree of accuracy the approximate position where the break along the eastern coast of the United States was to be expected. This coincided remarkably well with the observed Fall Line and emphasizes the usefulness of the techniques outlined above.

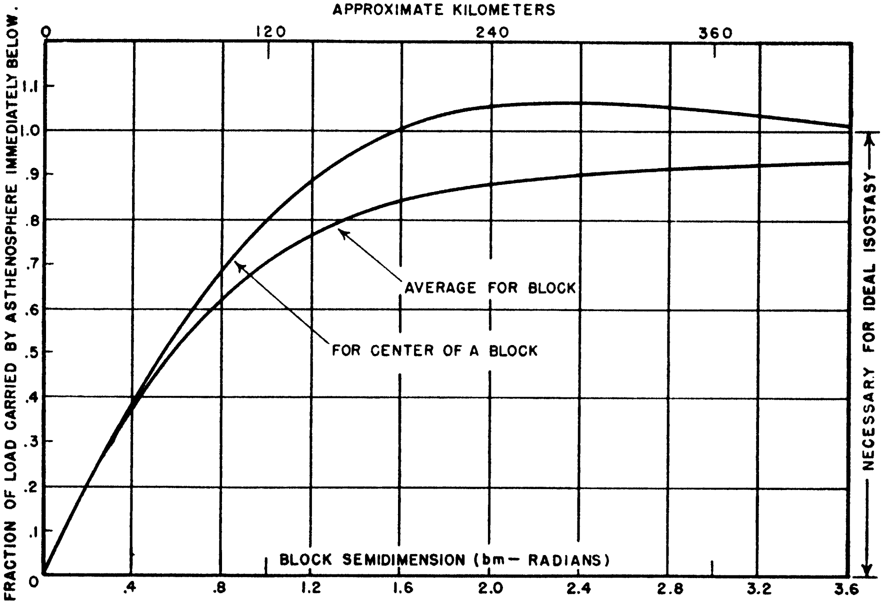

FIG. 3.—The fraction of a long block load carried by the asthenosphere immediately below

COMPARISON OF VARIOUS TYPES OF LITHOSPHERIC EQUILIBRIUM

The significance of the foregoing work, summarizing the mechanical behavior of a strong elastic lithosphere, is best understood by comparing the new assumptions with earlier ones.

Three fundamental hypotheses may be made concerning the lithosphere. It may be assumed to be (*a*) perfectly rigid, (*b*) plastic, or (*c*) elastic. The implications of conditions *a* and *b* are readily formulated in quantitative terms and have been widely employed in the calculation of the Bouguer and the isostatic anomalies, while condition *c* is the assumption made in this communication. Comparison of the three assumptions is summarized in table 1. It should be noted that the principal term of the gravity anomaly for an elastic lithosphere reduces to zero when $dh = d_0y$, corresponding to the condition for ideal isostasy, and to $2\pi\gamma dh$ when the crust is perfectly rigid and $y = 0$. Thus the elastic hypothesis leading to the principle of isobaric equilibrium, as here outlined, embraces the others as special cases.

TABLE 1

COMPARISON OF VARIOUS TYPES OF LITHOSPHERIC EQUILIBRIUM

Assumed Character of Lithosphere	Sufficient Approximation for Actual Earth when Load Is:	Resulting Character of Equilibrium	Lithospheric Constant	Type of Anomaly	Principal Term of Gravity Anomaly*
Rigid........	Small and concentrated	Mechanical	Zero	Bouguer	$2\pi\gamma dh$
Plastic........	Uniform and larger than 330 km.	Hydrostatic	Infinite	Isostatic	Zero
Elastic........	Unrestricted	Mechanical and hydrostatic	8.4×10^{-8} cm^{-1} (Gunn)	Isobaric	$2\pi\gamma(dh-d_0y)$

* In these quantities, h is the height of an impressed load having a density d, y is the downward deformation of the neutral axis of the lithosphere, d_0 is the density of the underlying magma, and γ is the gravitational constant.

REMARKS

The foregoing outline of the basic principles of isobaric equilibrium emphasizes the limited applicability of the principle of isostasy and attempts to achieve a general understanding of the structure of the crust and its behavior under geological loadings of various types. The investigation is more than a study of a special type of regional isostasy based on well-established mechanical principles because it permits one to forecast, for the first time, the magnitude and distribution of gravity anomalies from structural data and, conversely, to deduce the structure from gravity-anomaly data. Investigations outlined in this paper show that gravity anomalies do not, in general, imply an unstable geological condition but indicate, rather, that part of the stabilizing system of forces arises from the strength of the lithosphere. The outlined principles are of the utmost importance in understanding the basic mechanisms responsible for mountain-building. In fact, this investigation was an outgrowth of the two dilemmas of mountain-building that result from applying the classical isostatic theory. These two dilemmas were (*a*) the impossibility, according to isostatic theory, of laying down prisms of sediments of adequate thickness and (*b*) accounting for the very rapid rise of geosynclines after an extremely long period of development and equilibrium. A strong elastic crust having the properties outlined above is necessary to resolve these inconsistencies (Gunn, 1937).

The foregoing conclusions also modify in an important manner some of the basic ideas concerning the "roots of mountains." If the conditions of ideal isostasy prevailed, actual mountainous "roots" would always be formed, and they would extend downward for a considerable distance. But, actually, because of the great strength of the lithosphere and its elastic characteristics, an appreciable share of the weight of most mountains of medium size is carried by the lithosphere and is spread out over a distance approximately 250 km. each side of the load. Thus the root and the compensation is to a considerable extent not under the mountain itself but spread out to the sides. The resultant roots under the mountains consequently are relatively much shallower but much more extensive than one would estimate, were the region in isostatic equilibrium.

The effect of a strong lithosphere in spreading the ultimate support for loads having a limited area over a considerably greater area emphasizes the artificiality of the isostatic hypothesis. In the reduction of isostatic data, the assumption is made that the support or compensation is immediately below the load. Although this assumption simplifies the computations, it may be seen from figure 2 that it is not justified and is hardly descriptive of the actual geometrical configuration. It is evident, therefore, that gravity anomalies derived from isostatic reductions are commonly unreliable and misleading except when applied to relatively large areas showing little deformation. The free-air gravity anomalies are much to be preferred in discussing geological structure because they more nearly represent appropriately corrected observed data and are readily computed. Any proper topographical corrections must be made with full allowance for the fact that the support for any given earth feature is distributed over a considerable area.

It is recognized that the introduction of a strong elastic lithosphere between a given load and its ultimate hydrostatic support spreads the load and complicates the laws of crustal equilibrium. However, the principles of "isobaric equilibrium" well describe the actual mechanical balance of the lithosphere and are unusually fruitful in describing various geological phenomena in terms of measurable gravitational variations. This extension of isostasy will, therefore, add notably to the interpretations of structural geology.

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A REVIEW OF TECTONIC RELIEF IN AUSTRALIA[1]

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ABSTRACT

Süssmilch and others have maintained for many years that tectonic scarps in eastern Australia, mainly fault scarps, are prominent in New South Wales, and more recently they have described the eastern ranges of Queensland as made up of elongated fault-block features. Alternative explanations recognize a prevalence of multicycle relief forms in New South Wales and attribute the features of the Queensland ranges to subsequent erosion on an ancient structure.

In Victoria, South Australia, and parts of Western Australia, large block-faulted relief forms are recognized; but in the first-named state some scarps formerly regarded as tectonic have been more recently relegated to the fault-line scarp category.

In Tasmania relief is dominated by fault scarps which are in a sense also structural escarpments; and survival of these in a single cycle from a Lower Miocene tectogenesis is claimed on the strength of paleobotanical evidence.

INTRODUCTION

Lack of agreement as to what are and what are not fault scarps in Australia and the unsettled condition of opinion regarding tectonic relief in general in the eastern part of the continent point to a need for further investigation of some of the major features of Australian relief. Various methods of investigation of major relief have been employed, ranging from one which confines attention to underground structure to that based on recognition of the dislocation of former land surfaces tentatively restored for the purpose; but the features of few landscape scarps themselves have been fully investigated or described.

A theory of prevalent tectonic relief in eastern Australia was advanced long ago (Andrews, 1910), and it is now generally agreed (though possibly with some dissentients) that some of the major features had such an origin. With regard, however, to the extent to which tectonic relief prevails, there are distinct schools of thought.

NEW SOUTH WALES

Süssmilch (1910–1937) has maintained for many years that prominent tectonic scarps, mainly fault scarps, are numerous in New South Wales. On the other hand, however, some high scarps formerly thought to be tectonic have been explained as resulting entirely from long-continued selective erosion of a heterogeneous terrain. It is therefore suspected that such an explanation might be found applicable to other scarps also, with the result that all scarps for which a tectonic origin has been suggested are regarded in some quarters with the strongest suspicion.

It is indeed desirable that "the sure methods of stratigraphy . . . and of detailed geological mapping" recommended by some Australian geologists should be applied to the problem, and all relevant geological data must be considered. It is a fact, however, that there is a considerable lag in geomorphic description and, in the absence of this, geological methods of investigation that do not amount to exposure of faults and reliable inferences as to date of faulting lead in many cases to inconclusive results. Some work-

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ers, notably Süssmilch, have placed confidence in doubtful evidence of dislocation of the basalt sheets (of early Pliocene age, according to Süssmilch [1937]) of subaerial origin, which are widespread in eastern Australia. Inherent in this view of large-scale landscape faulting is the conception of nearly perfect peneplanation over a very large area, making the Tertiary "great Australian," or "Eastralian," peneplain.

Faith in this conception has been shaken, however, by some investigations (such as that of Voisey [1942]). These show (1) that the land surface on which basalt lies is, in some districts, one of rather strong relief and (2) that this relief was ancient when buried, being of a residual nature above the general level of the Tertiary peneplain in some extensive areas characterized by the presence of large bodies of resistant rock in the terrain. These areas include not only the New England Plateau of northern New South Wales, in which ancient relief has long been recognized by Andrews, but also the Southern Tablelands (Craft, 1932, 1933*a*, *b*). Here, it is pointed out, basalt capping is found at very different altitudes, but this may not be interpreted as a result of dislocation of a once continuous sheet covering a peneplain. "On the contrary they [the basalts of Monaro and Kiandra] are associated with surfaces which bore the same relation to one another as they do now."

Throughout southern New South Wales and bordering the New England Plateau of New South Wales more or less prominent scarps have been diagnosed as of tectonic origin by some authors. Undoubtedly, the early to middle Tertiary cycle of erosion was preceded in eastern Australia by a long period of earlier subaerial denudation. The region has been above the sea since late Paleozoic times, and it seems that in the Tertiary cycle only the final touches were given to the eastern Australian peneplain, where it became widely developed over areas underlain by weak rocks, being there remade from a more ancient peneplain. It seems impossible to reject the belief that for some reason the tempo of denudation was very slow in this region, thus allowing the *survival of much relief in hard-rock districts*. This was not mountainous relief, which is necessarily short-lived, but there were very extensive residual reliefs on large bodies of resistant rock, which have been only very gradually subdued in successive cycles during eons of erosion. Such an explanation of the landforms of the New England district has been generally accepted since it was offered many years ago by Andrews; but its application to the extensive plateaus of southern New South Wales has been made gradually and cautiously by various workers, notably Browne (1929, 1944) and Craft (1932, 1933*a*, *b*). More recently, Maze (1944) also has recognized down-stepping peneplain benches west of the Blue Mountains.

One important conclusion reached by Craft is that much of the relief of eastern New South Wales has survived a denudation following post-Mesozoic and even pre-Mesozoic updomings. "The highlands thus formed have been subjected to continuous invasion," however, "and incomplete peneplains have been eroded, . . . which lead to higher cores of gently concave aspect."

Notwithstanding the discovery that large parts of the area of New South Wales were diversified by considerable relief prior to the upheavals which are attributed to late Tertiary crustal unrest—the almost-orogeny of Andrews' Kosciusko epoch—the formation of tectonic surface features as a result of diverse upheavals of this date cannot be doubted. However, the magnitude and, in particular, the abruptness of the tectonic relief have been perhaps overestimated by Andrews (1910, 1934), Taylor (1910, 1911, 1918, 1923), and Süssmilch (1910, 1911, 1914), who have been convinced of the tectonic origin of most of the large landscape features. Craft, on the other hand, relegates tectonic relief to comparative unimportance. He recognizes no conspicuous scarps of this origin, though he allows for differential upheaval in some upwarped areas. "Viewed as a whole," he writes, "the region has been subjected to a gentle post-basaltic uplift of 2000 feet on the median line of the highlands, with relatively depressed portions."

That the true explanation lies between the extreme views that have been advanced—that certain scarps, as well as extensive broad domings, have been developed and still remain features of the landscape—is shown by a description recently published of a stepped or splintered fault scarp descending eastward from the highest (or Kosciusko) plateau remnant in southern New South Wales (Browne, Dulhunty, and Maze, 1944). There is also the unchallenged authority of David (1908) for the occurrence of fault scarps in that locality. It appears, however, that some major scarps which have been diagnosed as tectonic in southern New South Wales must now be considered the unfinished work of secular subaerial denudation on a heterogeneous terrain of rather ancient rocks.

Perhaps the straight Cullarin scarp, or rather the diagnosis of this scarp by Taylor (1907) as a fault scarp, will stand the test of a long-delayed full discussion which it deserves. This scarp, about 40 miles long but of moderate height (about 500 feet), is the western boundary of the remarkable Lake George Basin of interior drainage on the broad crest (near Canberra) of the main divide between eastern and Murray Basin drainage to the ocean. This scarp defines the line of one of "two great faults" postulated by Taylor (1910) to account for prominent lineaments of the landscape of the Federal Territory at Canberra, though on the second of these lines, that followed by the Murrumbidgee River for a long distance, faulting either modern or ancient seems to be a very remote possibility according to the geological work of Browne (1944).

Discovery of the Cullarin scarp inspired Taylor to produce a geomorphic description more detailed perhaps than any other to be found in Australian literature; and probably this is one of the best authenticated of Australian fault scarps. After describing the sprawling spurs that descend between broad valleys into the alluviated Lake George plain from the range east of it, Taylor continues:

> Reverting to the western shore, we are struck by the absolute dissimilarity. Standing at the level of the lake we seem to be confronted by a great wall extending northwards for over 20 miles from the Molonglo Plain. No broad valley breaks its continuity. Indeed . . . it seems unbroken by any definite gap, while for a large part it presents a steep face 500 feet high to the lake. From the eastern shore, however, one is able to see a well-defined gap about halfway along the western shore, where the old Southern road crossed the Cullarin Range . . . Geary's Gap. . . . Less than two miles south of Geary's Gap a stream, Grove Creek, rushes down to the lake. Contrast its course with that of Taylor's Creek, on the eastern shore. The latter stream flows through a broad valley a mile or two wide, and shows the even grade of a long established river. The Grove Creek is barely a mile long, yet descends nearly 300 feet. Its course is interrupted by falls 25 feet high, and finally it emerges from a gorge or miniature canyon with steep sides 200 feet high. Evidently it is a stream which has barely reached the youthful stage.

He finds also that the Molonglo River enters the scarp farther south and crosses an upheaved block west of it by way of an antecedent gorge. Geary's Gap may be a similar gorge which has become a wind gap as a result of defeat of the stream which had followed a course through it until it was ponded in the Lake George Basin by acceleration of the upheaval. Similar features furnish good evidence of block faulting in other regions.

Acceptance of Taylor's description and diagnosis of the Cullarin scarp as a fresh, or very youthful, fault scarp perhaps distinguishes it from other Australian tectonic scarps, most of which are rather the worse for wear. Either the dislocation on the Cullarin line took place more recently than most of the others, or circumstances have combined to slow down the erosional degradation of this scarp. The latter may well be the correct explanation, for no erosion by a through river seems to have taken place in the Lake George Basin which could affect the form of the scarp. The endeiric drainage of the basin is unique in eastern New South Wales, and the peculiarity is attributable to a combination of climatic conditions with accidental disposition of blocks and the pattern of the land surface consequent on deformation. Remoteness from rivers and immunity from dissection except by short, consequent streams may allow a low scarp to retain its primitive form almost indefinitely. Arid erosion, moreover, favors the perpetuation of steep slopes, though these may retreat without appreciable change of form.

It is of some interest to turn from the Cullarin scarp to examine reasons which led both Taylor (1910) and Süssmilch (1910) to regard as tectonic another long scarp in the Canberra district, that which descends to the meridional reach of the Murrumbidgee River from the highland to the west of it, although Browne's geological work in the district does not support the diagnosis and Craft (1933*a*) regards the scarp as a feature surviving from Tertiary and pre-Tertiary relief. For identification of a fault scarp on this Murrumbidgee line Süssmilch has relied on a postulated dislocation of a formerly continuous peneplain, which he claims was very widespread and low-lying throughout eastern Australia in mid-Tertiary times. The dislocation, assumed to take place at the time of the generally recognized late Tertiary or end-Tertiary upheavals, would make a scarp 2,500 feet high. Taylor's diagnosis of faulting is also based in part on an assumption that it is necessary to postulate such dislocation of the landscape to account for a rectilinear scarp along the left bank of the Murrumbidgee River. However, it must be admitted that, apart from a regional break in the topography, such details of landscape as are shown by Taylor's maps and the photograph of his relief map of the Federal Capital Territory do not suggest a fault in the same vivid way as on the Cullarin line.

In explanation of the scarped valley-side profile, the fact is significant that streams tributary to the Murrumbidgee River flow, probably because of adjustment to some ancient structure, parallel and close to the main river for long distances and thus rob it of the accustomed quota of valley-side drainage. Where any linear cliff is deprived in this way of the water necessary to feed even rivulets, its dissection is necessarily greatly retarded. A valley side thus assumes and long retains a wall-like form resembling that of a young fault scarp.

Some generalized accounts of the physical features of New South Wales imply that, apart from those which may be attributed to marine cliffing, there are distinct scarps along the coastal margins of high plateau areas which must be tectonic. According to Craft (1933*b*), however, who bases his opinion on a measured gentleness of regional slopes, it is unnecessary to postulate that there was initially any abrupt seaward descent or variation from very gentle regional monoclinal warping, which has, moreover, merely steepened a regional slope inherited from distant prebasalt times. "The plateau edges are primarily erosional features," and "when the rocks of the landscape are more homogeneous the streams take on a symmetrical character" on the seaward marginal slopes. As Craft points out, however, "portions of relatively undissected plateau come near the coast at intervals throughout its length." At such places it is possible that the initial seaward descent has merited description as tectonic scarps. This is suggested by some statements of Andrews (1914), such as: "At Illawarra and still farther southwards the plateau advances almost to the shoreline"; but, actually, where the seaward face of the inland plateau becomes a high and bold scarp, this is a structural escarpment. That of the Illawarra coast, which Andrews cites, has resistant Triassic sandstone as cap rock, and the coastal profile of which it forms a part may have been formed by the processes of erosion from a gently inclined monoclinal slope just as well as from a fault scarp.

A very gentle seaward tilting of a surface already sloping seaward at the end of the Tertiary era has been sufficient, according to Craft, for the rejuvenation of the valleys of coastal rivers in both northern and southern New South Wales, thus causing formation of falls and the excavation of gorges, commonly 1,500 feet deep and in some cases deeper. In northern New South Wales (New England) the presence of a basalt capping on the plateau at the headwaters of some of the coastal rivers has checked their headward erosion and steepened their gradients at a zone of escarpments; and it is questionable whether any of the basalt escarpments at the margin of the high plateau, however imposing, are derived from true tectonic scarps. An example of such is the escarpment at Guy Fawkes, overlooking deeply cut headwater valleys of the Bellingen and Nambucca rivers, "the view from which is probably unexcelled in New South Wales, or even in Australia" (Andrews, 1914). According to Andrews, this feature may "be explained either by erosive or by tectonic forces." He adds: "Similarly for the great eastern scarp overlooking the South Arm of the Clarence." Though (at another place) he views the marginal feature emphasized by these escarpments as "strongly suggestive of an important fold or fault," it appears that no positive evidence of the presence of a tectonic scarp has come to light. It is significant, on the other hand, that in a part of the seaward slope from New England which Craft selects as typical, the basin of the Macleay River, he finds no strong scarp in the present-day form of the surface, not even an escarpment at the margin of the basalt capping.

Süssmilch has not attempted to recognize tectonic features in either the seaward or the inland margins of the plateaus, merely remarking (1937) that "since their uplift the tablelands have suffered considerable dissection, particularly along their eastern and western margins." Benson (1917) has noted, however, that, although "the plateaux of New England sink to the west and pass out into the plains," one high salient or buttress forming the Nandewar Mountains is bounded on the west (on the authority of Jensen [1907]) by a tectonic scarp.

One high and prominent scarp in New South Wales has been given some recent attention. This scarp, on a line recognized even by Craft as a fault ("Hunter boundary fault"), forms part of the southern margin of the New England plateaus. There is here a steep descent to the low-lying Hunter River Valley, a tectonic sag, much enlarged by erosion, which is continued westward in the Goulburn Valley and the major geographical feature called by Taylor the Cassilis "geocol." This sag separates the northern and middle members of the three broad upswellings comprising the New South Wales portion of the eastern "cordillera," or Main Divide, of Australia, which are described by Craft (1933*b*) as "three low domes," although, according to him, the Main Divide as a whole is "of essentially non-tectonic origin." Süssmilch (1940) has described parts of this scarped descent from the Barrington salient of the New England plateaus westward and southward to the Goulburn-Hunter Valley, where fault scarps had been earlier diagnosed by Andrews (1910).

According to the description given by Süssmilch, these slopes contrast strongly not only with the subdued landforms of an extensive bench at their foot, which is regarded as part of a maturely dissected peneplain, but also with the small relief of a plateau above, though exact correlation in the plateau and bench of dislocated parts of a formerly continuous land surface has not been made. About 60 square miles of highland plateau survive near the western edge of the Barrington salient at an altitude of about 5,000 feet. The plateau consists superficially of basalt, which is found to be a capping 500–800 feet thick over peneplaned granite and other rocks. West of a marginal scarp, presumably tectonic, the base of the basalt is 1,200–1,500 feet lower, indicating a lowering in that direction along a "post-basaltic fault or a monoclinal fold." Topographically the scarp form is not very clearly marked, but rivers which presumably originated in courses consequent on its slope are superposed in gorges across strike ridges of Devonian and Carboniferous bedrock. Farther west at a lower level a second west-facing scarp of similar height is regarded as another fault scarp in a stepped descent. This one is related to the known Wingen fault, which brings Permian and Triassic strata on the west side against the folded Carboniferous. The displacement here, as measured on dislocated basalt capping, is 550 feet; but the scarp on this line remains a strongly marked topographic feature only where parts of it take on the character of a structural escarpment, owing to the capping of basalt. The well-dissected condition of these scarps indicates that the Kosciusko uplifts to which they are attributed were not very recent.

Along the southern margin of the Barrington salient, or southwesterly projecting buttress of the New England plateaus, there is a scarp feature which remains more strongly marked topographically and is described by Süssmilch as "a magnificent scarp trending almost due east and west and giving a drop of more than 3000 feet from the higher to the lower tableland." It "displays all the characteristic features [not, however, enumerated] of a true fault scarp, and was diagnosed as such by E. C. Andrews in 1910." The fact that the scarp appears to be entirely in basalt is explained by Süssmilch as due to a misleading effect produced by the presence of much basalt talus on it, which entirely masks a bedrock consisting presumably of ancient rocks. No alternative hypothesis of origin of a structural escarpment (capped by the basalt) at this line without faulting seems to offer a satisfactory explanation of the feature.

The middle member of the eastern Australian group of high plateaus, the Blue Mountains—or Blue Mountain Plateau or Tableland, as it is better termed—presents the least ambiguous tectonic feature in the whole region, an enormous monoclinal flexure of the land surface, which descends from high plateau summit levels (in places about 4,000 feet) nearly to sea level a short distance west of Sydney. This monocline and its local steepening northward into a conspicuous scarp were noted by David (1902) and Andrews (1903). The steeper scarp portion is described by Taylor (1918, 1923) as monoclinal and by others as a fault scarp. Such features, when they have undergone dissection, are not readily differentiated.

A brief general account of the landscape of the Blue Mountains as a whole has been given by Taylor (1918) as follows:

> The Blue Mountains consist of an ancient peneplain, which is a relic of the much greater land surface of early Tertiary times. The eastern portion has sunk beneath the sea, while the central portion has been buckled up to the extent of several thousand feet. . . . The western slopes . . . merge gradually into the Western Plains, and the abrupt scarps of the eastern side are not present there.

Taylor considers the peneplain to have been similar, prior to its upheaval, to that which survives far down the western slope of the arch, where the "Cobar peneplain" is surmounted by scattered low monadnocks with maximum relief of 500 feet.

According to Craft (1933*b*) also:

> A general relief of 500 feet may be ascribed to the Blue Mountains portion of the pre-basalt surface, and a possible altitude of 1000 feet above sea-level, with such ancient residuals as Mount Lambie rising almost to 2000 feet. . . . The surface has been deformed. . . . There has been tilting or faulting towards the low country. . . .
>
> At the time of the basaltic extrusion the Central Plateau or Blue Mountains formed a low region between the higher masses. Its general elevation probably did not exceed 1000 feet, with a maximum of 2000 feet, whilst the higher mass of New England rose to 3500 feet, and the Monaro-Kosciusko region to 4500 or 5000 feet as a maximum. There appears to have been little subsequent deformation of the higher regions, but the Central Plateau may have been warped considerably.
>
> . . . The most complete peneplanation recorded in the history of existing landforms is in the region of the Blue Mountains, where monadnocks are comparatively insignificant points.

The Central Tableland as defined by Süssmilch (1914) includes, besides the Blue Mountains, other considerable areas from 2,000 to 3,000 feet in altitude, which he has regarded as distinct blocks bounded and separated from one another by tectonic scarps. These scarps he has not described, however, except as "abrupt changes in elevation," which are likened to the steeper part of the Blue Mountains monocline. The southern boundary of the Blue Mountains unit he has termed a "magnificent fault escarpment," but the correctness of this identification of a fault scarp is perhaps questionable; for Craft (1933*b*) mentions no appearance of faulting of the land surface at or south of this line and, indeed, describes the southerly descent from the Blue Mountains as a warped surface.

From the foregoing remarks regarding Kosciusko upheavals and dislocations and post-Kosciusko dissection in New South Wales, it will be gathered that the tectonic features of eastern Australia are of rather ancient origin and are for the most part well dissected. A post-Pliocene date seems to be out of the question for these upheavals; but there is no reason, on the other hand, for assuming that the origin of the surviving tectonic features may be dated earlier than Pliocene, notwithstanding the apparently necessary postulate of a slow tempo of denudation in this region.

QUEENSLAND

A recently published summary by Süssmilch (1938) of geomorphic evidence strongly favors recognition of numerous tectonic scarps, and fault scarps in particular, in the eastern coastal districts of Queensland. He writes:

> In most parts of eastern Queensland very little geological mapping has been carried out; consequently definite geological proof of faulting is not available for many of the supposed fault scarps. These particular scarps are, however, so similar in their physiographical features to those for which geological proof is available that it is quite reasonable to assume that all of them have had a similar origin.

He lists several categories of scarps which he regards as fault scarps and continues:

> It may be true that some of the scarps are due to warping rather than faulting, but this would not alter the fact that they are tectonic scarps and not erosional scarps. If this conclusion is correct, then the coast ranges and chains of continental islands are fault blocks, the corridors, both land and marine, are rift valleys, while the coastal plains and broad lowland regions are still-stand areas—i.e., parts of the late-Tertiary land surface which lagged behind during the general uplift.

Prior to the publication of this contribution by Süssmilch and that of Stanley (1927), who has admirably described a portion of the Queensland coastal area, generally similar conclusions had been reached, though stated in broader terms, by Andrews (1910) and by David (1914). Some Queensland geologists, however, notably Marks (1924) and Bryan (1928, 1930), have scouted the idea of the existence of any tectonic relief, mainly on the ground that the trend of elongation of coastal features, ridges, and corridors finds a sufficient explanation in the grain of the folded Paleozoic terrain. Thus Bryan in a broad generalization attributes all elongation and alignment of features that lie more or less parallel to the coast to dominance of the north-northwesterly strike of the Paleozoic structure. The river courses, together with corridors which other observers point out are not and probably never have been occupied by through rivers, are explained as "subsequent"—i.e., of an origin attributable solely to selective erosion; and what is true of valleys must be true likewise of interfluvial highland ridges.

One of the main arguments employed by Süssmilch in Queensland, as in New South Wales, is based on recognition of dislocated parts of a formerly continuous peneplain under the offshore shelf, on the floors of low-lying coastal strips and corridors of small relief, and on the crests of highland blocks in the coastal ranges. No plateau (peneplain) remnants are recognized on the coastal ranges by Bryan, however, although Andrews (1910), Daneš (1911), and others have there thought to identify parts of a well-worn surface which has commonly been regarded as a peneplain of Tertiary development, presumably continuous with that of New South Wales. On such a surface in Queensland, according to Stanley (1927), "dissection had reached maturity, . . . *but peneplanation was not nearly complete.*"

Regarded as an important "datum" by Bryan (1939) are red earths and laterite of a series of which the origin is attributed to a moist Pliocene epoch immediately preceding the Kosciusko uplifts of New South Wales. Though found at many places in southeastern Queensland on a level surface, and on basalt covering such a surface, this soil formation has been reported only at low altitudes on straths of an imperfect partial peneplain which has failed to truncate the more resistant rocks of the region. The discovery of the red-earth datum has not yet afforded a key to the geomorphic history of Queensland; but, if the date attributed to it is acceptable, it seems to push back the date of any recognizable higher-level surface of advanced denudation there may be to a remote past. This is in conformity with the theory of slow tempo of Australian geomorphic development. Yet the uncertainty of the date of origin of such a surface, if it is real, does not affect the value of inferences based on its presence at apparently dislocated levels.

Among the most striking features known as major landscape forms in Queensland are the so-called "corridors" (an empirical term first employed by Jardine [1925] for them). According to Bryan's interpretation, they are no more than valleys formed at some time by subsequent rivers; but, according to the tectonic interpretation, they are grabens. They are elongated parallel to the coast, with either slightly emergent or submerged or buried floors. Intercorridor horsts, corridors, belts of outlying islands, and a hypothetical bedrock bench generally believed to underlie the very broad continental shelf and Great Barrier Reef at a shallow depth are elongate features aligned parallel to a coast whose initial form, according to the tectonic explanation, was caused by rather recent earth movements.

The evidence of marginal and block faulting is found for the most part in steep regional slopes that separate "tablelands," which surmount the highest ridges in Queensland (all in the coastal belt), either from benches which are apparently step-faulted or from nearly level corridor floors, "coastal plains," and the undersea surface of the continental shelf, which is regarded as down-faulted. Süssmilch (1938) has drawn attention to these by sketching generalized profiles of the landscape, which are designed also to show plateau relief on the ranges.

Beyond this, the geomorphic evidence as to the nature of the scarps is only vaguely known. Proponents of the theory of tectonic relief have not, as a rule, given much attention to description and discussion of the features or to the stage of dissection of the fault scarps which they have diagnosed. That this dissection is fully mature may be inferred from such statements as "both groups of scarps [there are scarps transverse as well as parallel to the longitudinal topographic elongation or trend] display a similar amount of dissection" (Süssmilch, 1938). A similar inference may be drawn from the statement that "differential erosion has played an important part in the dissection of the tablelands and horsts since their uplift," for this allows time for, and implies, advanced dissection of the scarps that bound these features. If these tectonic scarps were still young, there could be but little dissection as yet of upland surfaces behind them. Yet differential erosion is reported to have opened valleys right through some horsts along belts of weaker formations which separate granite massifs.

In addition to the longitudinal scarps attributed to faults bounding horsts, grabens, etc., some are mentioned which are "transverse to these trends." As these latter are transverse also to the structural grain of the terrain, they are perhaps more plausibly related to faults, although whether these be ancient or relatively recent dislocations is another matter. It cannot be said with certainty as yet that differentiation has been clearly made in Queensland between fault scarps and fault-line scarps; but the balance of probability seems in favor of the tectonic origin that has been confidently ascribed by various authors to scarps which appear to bound horsts and grabens in the coastal belt.

Prior to the publication of Süssmilch's generalization, Stanley (1927) vigorously advocated the tectonic theory, which he applied to a typical portion of the coastal belt, describing it as the "flexed and faulted-down margin of the Pliocene continent." Describing Whitsunday Island, one of a group rising from the submerged shelf which David (1911) long ago referred to as "almost certainly horsts among a network of faults," Stanley wrote:

> Observe that there is here no question of differential erosion. The physiographic units are constituted of many different types of rocks, and they cut across the geological boundaries. In the island festoons both granites and Paleozoic volcanics figure equally, and the same is true of the coastal ranges. Important as rock structure and relative resistance to weathering may be in determining topographic forms, it does not seem feasible to consider that it has acted as the broad control which is required by the systematic arrangement of the units.

Some Queensland rivers are very strongly rejuvenated in their lower courses, and this fact has often been cited as proof of the presence of either a coastal fault or a strong flexure of the surface. Taylor (1911) and Daneš (1912) have each described the Barron River and its falls as an example. This river, Taylor wrote,

> leaves the highlands . . . by falls 600 feet high. This gorge cannot be due to ordinary uplift, for the coral reefs in the immediate vicinity show practically no sign of elevation. But it is quite probable that the conditions which obtain in the south-east of the continent—namely, blocks of the earth's crust bounded by north-south faults—have also determined the topography here. We may imagine (as Dr. Daneš suggested) a Tertiary extension to the east for several hundred miles with a normal consequent drainage. The greater portion of this is faulted and subsides gradually, and is covered by the coral reefs of the Great Barrier. Only the headwaters of these rivers remain, and they are now in a much more favourable position than before the subsidence as regards rapid erosion. . . . Faulting occurs. . . . The river is betrunked in a very drastic fashion. The short coastal Barron has now to drop 1000 feet in ten miles instead of in several hundred miles, and hence cuts out a gorge very rapidly.

VICTORIA

In eastern Victoria (the Eastern Highlands) the high-plateau country of the Kosciusko district of southern New South Wales has its southerly continuation. Nothing is definitely known in this region of any tectonic scarps of Kosciusko or earlier upheaval, although Hills (1934) notes that differential movements have undoubtedly taken place. Conspicuous scarps, however, both here and in the Grampians (Western Highlands of Victoria) have been shown to be due to selective erosion working on a heterogeneous terrain (Hills, 1936). In the southern part of Victoria, on the other hand, there is ample evidence of differential movements which may have been contemporaneous with Kosciusko uplifts, and fault scarps are recognized (Hills, 1940).

According to the findings of Hills (1934, 1940), the geological history of Victoria is different in some important respects from that generally accepted for New South Wales. The most ancient land surface in Victoria is preserved in some small residuals of high dissected plateau on areas of very resistant terrain; and this oldest surviving peneplain is attributed to a cycle interrupted before the end of the Cretaceous period. A later, lower-level peneplain was extensively developed at the expense of the dwindling Cretaceous surface prior to the extrusion of extensive cappings of basalt, presumably in the Oligocene period. The Oligocene land surface—a peneplain with not inconsiderable monadnocks still standing above it—is known largely as a fossil surface because of its preservation under the basalt. Following this, middle Tertiary erosion planed off considerable areas which were later submerged in marginal districts and buried under marine strata; and it is subsequent to this (in the Pliocene period) that the upheavals are dated which have made the more prominent features of Victorian relief (the Eastern Highlands and the Central Highlands). The major relief forms which resulted from these middle to late Pliocene movements still survive; but the movements were, according to Hills, of a pre-Kosciusko date. In Gippsland (southeastern Victoria) the land surface was dislocated by movements of probably later, i.e., very late Pliocene, date (Kosciusko?).

The lack of agreement between events and their sequence in New South Wales and Victoria is perhaps only apparent, for it is possible that dates attributed to major earth movements in Victoria will be found to be applicable over a wider region. The date assigned by Hills to Kosciusko uplifts (in the most restricted sense) is late Pliocene to early Pleistocene, at which time he believes that movement was renewed in the Central Highlands, while at the same time new ranges made their appearance in southernmost Victoria. There, especially in the South Gippsland highlands and in the Otway Ranges farther west, a tectonic relief of very late origin is recognized. One of the best-marked scarps in South Gippsland is the maturely dissected Yarragon scarp, which faces north, overlooking an embayment of the East Gippsland Plains (a tectonic basin); but, in addition to this, several others are mapped by Hills along the northern and southern margins of the upheaved area.

Hills has attempted to distinguish with care between fault-line scarps (which are generally two-cycle features) and true fault scarps. Many supposed fault scarps in Victoria have been relegated to the former category; but (he states) "the South Gippsland highlands offer an excellent example of block mountains formed by Cainozoic faulting, and they constitute a well-defined physiographic region." At another place: "In south-central Victoria the district between the Rowsley fault and Selwyn's fault has been depressed, forming the Port Phillip sunkland. . . ." One of these faults, the Rowsley, regarded as of Pleistocene age, is marked in the southern part by a well-defined scarp, which Hills (1940) has selected as an example of a fault scarp dissected to maturity. It forms the eastern boundary (with 800-foot relief) of the Brisbane Range, a back-tilted block. Fenner (1918) in his description of this scarp has drawn attention to the fact that its topographic form complies with Davis' ideal fault scarp of the Wasatch variety in that it truncates indifferently the structures in an upheaved block and has a bahada in front of it. It is dissected by a number of short streams which are presumably consequent. "Deep gorges score the face of the high resistant scarp, and the streams continue on the low block in wandering shallow channels across the alluvial apron" (Fenner, 1918).

Hills has also figured the profile of the Otway Ranges uplift, farther to the southwest, as a tectonic feature limited by post-Pliocene monoclinal flexures steeper on the southeast (coastal) flank.

TASMANIA

Fault scarps have been pointed out by Carey (1947) in the Launceston district of Tasmania and are interpreted by him as ancient features, though in a good state of preservation. In his account of this district, which is accompanied by a set of comprehensive block diagrams in sequence showing paleogeography, strong tectonic relief is described and shown as having its origin in the Lower Miocene period, and no relief of any kind inherited from earlier times or developed later by earth movements is recognized. No indication has been found of even minor renewal of faulting subsequent to the Miocene paroxysm; and only very small movements of base-level unaccompanied by appreciable warping are believed to have taken place in later times. A very simple picture of tectonic and erosional history is thus presented.

The predeformational surface, dislocated in Lower Miocene times, was, according to this interpretation, a land surface in an advanced stage of senility after a resting period of ninety million years. "This peneplain reached a high degree of perfection. No protuberances suggestive of old monadnocks . . . have been recognized." The senile surface was crusted with laterite, especially where it crosses a very extensive sheet of Jurassic dolerite intrusive between Lower and Upper Permian strata.

The Miocene dislocation of this surface developed gentle folds broken by strong strike faults, so that the resulting relief was expressed mainly as a number of elongated, subparallel fault blocks disposed as steps on the flanks of a few broken anticlines.

The strong relief of early Miocene origin is depicted as surviving very extensively into the modern landscape. Some fault scarps which had rock of the same kind on each side, even granite, are shown as entirely obliterated, naturally enough, by thirty million years of erosion; but on the sites of ancient compound horst blocks with originally thick dolerite capping there are still ranges; and the scarps that bound these, though they might, if an alternative multicycle hypothesis were available to account for the relief, conform at least in some parts to the definition of fault-line scarps, must be interpreted as fault scarps which still survive with much of their initial relief, if the postulate of only one cycle of erosion can be maintained.

Until some other interpretation of the geological and geomorphic history is offered, the story told in Carey's diagrams of the extremely slow wastage of escarpments bounding dolerite caps on some of the ranges may be tentatively accepted as a working hypothesis; and these are shown by him as true fault scarps.

Assuming that no question can be raised as to the date of faulting, it is difficult, however, to accept unconditionally the one-cycle explanation of this landscape. For one thing, thirty million years have elapsed; and, for another, the fidelity to nature of certain parts of the land surface as depicted, or else its inferred history, may pardonably be doubted. Though some episodes of the history of the district have not been written, the landscape has been so graphically delineated in the diagrams (fig. 1) that the story thus unfolded is none the less open to criticism.

At the foot of Ben Nevis, a broad bench that has apparently been leveled by Miocene and post-Miocene erosion and is cut across granite at about 3,000 feet above base-level remains unac-

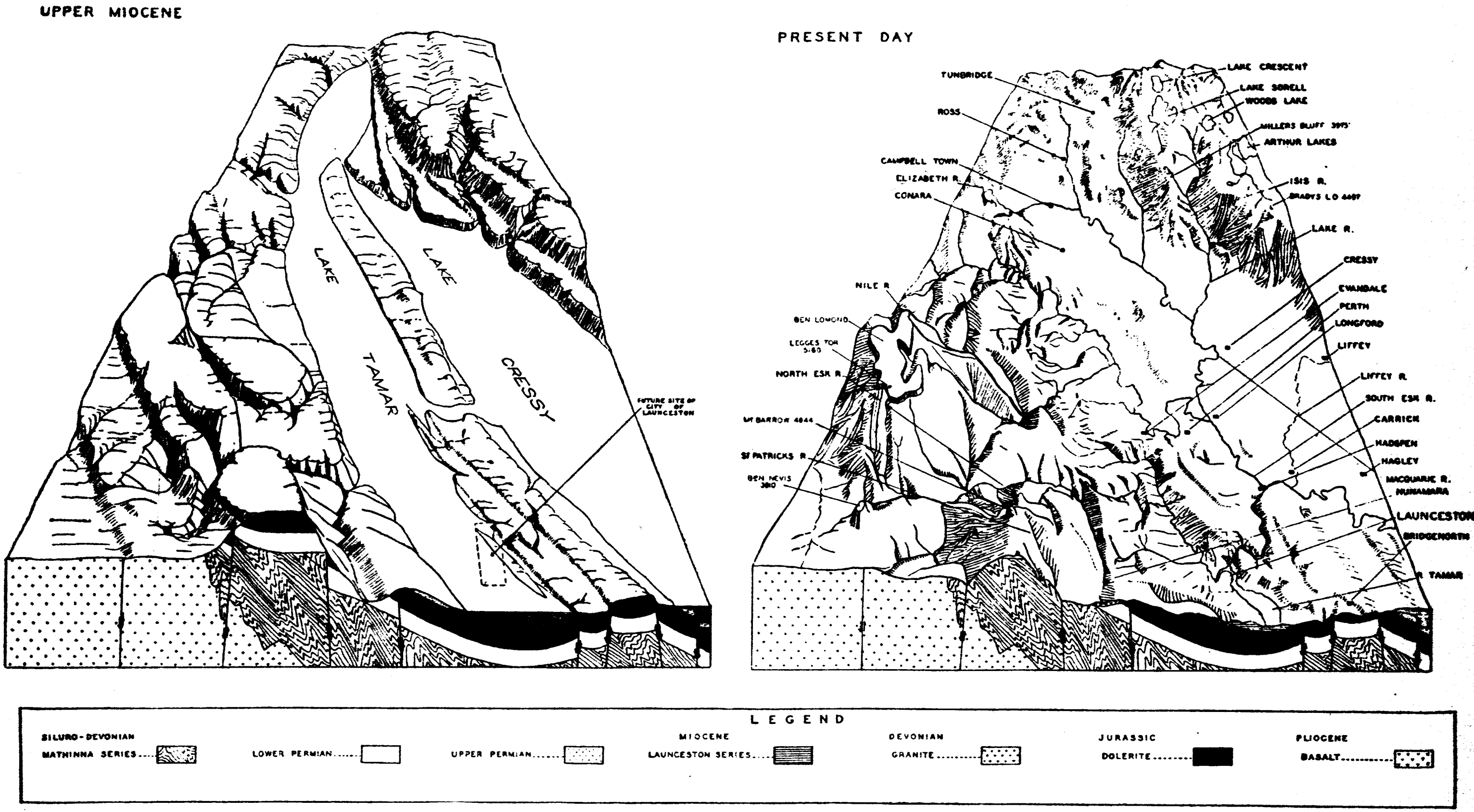

FIG. 1.—Block diagrams illustrating the geomorphic history of a rectangular block, 2,700 square miles in extent, in northeastern Tasmania. South at top. *At left:* **Upper Miocene** landscape, sequential after Lower Miocene tectogenesis. *At right:* The landscape of the present day (after S. Warren Carey).

counted for. It is at this high level of unexplained planation that denudation has obliterated a high fault scarp of granite. In contrast with this scarp degradation, in the southwestern part of the rectangular area figured, another fault scarp, which bounds the large pre-Miocene peneplain remnant forming the Great Western Tiers and is 4,000 feet high, retains its relief and primitive steepness, though exposed for an equal time to erosion, and has been worn back but a short distance. It is difficult to believe in the secular survival here and in Ben Lomond (5,000 feet high) of mesas of strong relief consisting of remnants of a dolerite sheet, which, though very thick, is weakened by its overlying the Permian mudstone at a high altitude. It would seem that the explanation given of these Tasmanian scarps must be considered as only tentative, or else that the date of the block faulting may be later than has been supposed.

The latter suggestion finds support in an account given by Lewis in 1939 of studies then in progress, which indicated his belief that the block movements responsible for producing the major features were of late Pliocene date, until which time the Tertiary peneplanation had continued, and that even in the Pleistocene Tasmania was still being broken into blocks along the lines which are now followed by consequent rivers. However, Lewis veered later to the opinion that the main landscape faulting was early Miocene, from which date, he wrote (1946), "a normal sequence of erosion has proceeded. This [current cycle] has only been broken by minor and local disturbances." He has assigned to early Miocene faulting the initiation of a very impressive scarp descending toward Hobart from Mount Wellington (4,000 feet), which lies immediately behind the city and is described as a compound faulted block thrust up in compression. Consequent dissection of the major scarp has produced "deep narrow gullies often encircled with low cliffs and ending in steep walls," and "the streams" are said to "have cut down rapidly to base-level, but have not yet had time to widen the valleys, thus indicating a fairly recent uplift" (Lewis, 1946). Lewis has, however, shown the high, bold front of Mount Wellington to be, like the great scarps of the Launceston district, in part a structural escarpment of the thick intrusive dolerite sheet overlying Permian strata which dominates the geology and landscape of Tasmania. Thus he invalidates the inference of "recent uplift," which is, in any case, inconsistent with the rest of the story of the geomorphic history.

The interpretation of at least the major scarps as tectonic rather than of fault-line erosional origin, being based on general geological relations, can scarcely be questioned; but when it comes to dating the earth movements responsible for making these features, the earlier guess of an end-Tertiary orogeny seems more acceptable to a geomorphologist than a middle Tertiary guess, unless much of the geomorphic story has been misunderstood or still remains untold. Adoption of the early Miocene date has made it possible to account for the presence of plant remains believed to be Miocene in lacustrine strata believed to have accumulated in hollows consequent on the great deformation.

SOUTH AUSTRALIA

In an account by Fenner (1930) of the landscape features of South Australia, he tells that "the Gulf and Torrens areas . . . represent major tectonic features that form the southern portion of . . . a great broken zone extending northwards across Australia." The boundaries of the main tectonic blocks (horsts and grabens) in the southern part of the region he shows on a map. He regards these as fault scarps and zones of step faulting and has described the stepped blocks, "particularly in the Flinders Ranges and in the eastern Mount Lofty Ranges," as "almost diagrammatic in appearance." The scarps bounding the Mount Lofty horst, especially those on the western side in the vicinity of Adelaide, have been described (Benson, 1911; Sprigg, 1945) and are well established by evaluation of both geological and topographic evidence.

In South Australia, unlike eastern Australia, large-scale tectonic (consequent) control of the existing drainage is recognized. Rivers of the past, so far as restored by paleogeographic study, followed different lines from the rivers of today; but, as a sequel to a late Tertiary fault dislocation of the landscape (as well as of a coastal strip of compound terrain with a cover of Tertiary strata), systems of streams have come into existence with centripetal arrangement in the Lake Eyre, Lake Torrens, and Lake Frome basins. These "are purely consequent upon and closely reflect the facts of the . . . ranges" (Fenner, 1930).

In an account published recently of the western slopes of the Mount Lofty Ranges, Sprigg (1945) has applied a statistical method of study based on topographic maps, which has enabled him to reconstruct generalized contours of a deformed surface, and has thus fixed the positions of scarps marginal to downstepping faulted benches. Sprigg has revived Benson's (1911) early fault-scarp explanation of these breaks in the continuity of an upland surface interpreted as a resurrected fossil peneplain. He does not favor a rather complex alternative hypothesis of faulting in two widely separated stages with an intervening peneplanation, as proposed by Fenner (1930) following an oral suggestion made by Douglas Johnson. According to this hypothesis the upland surface of the range consists in part of facets of the peneplain developed in the interval between two faulting upheavals. According to Sprigg's conclusion, on the other hand, the geomorphic cycle that was initiated by late Tertiary block faulting still continues. It "has removed the bulk of the [Miocene] overmass sediments as well as much of the early Tertiary erosion surface"—a senile surface truncating ancient rocks—which it has resurrected. The drainage of this complex block originated in consequent fashion on a dislocated and deformed surface of the overmass.

WESTERN AUSTRALIA

Only brief reference need be made to Teichert's recent suggestion (1947*a*) that all the deformation, including strong dislocation, that has been suffered not only by Tertiary but also by Mesozoic and late Paleozoic strata in Western Australia, has taken place in Pliocene times. Probably this is based on a misconception of the geomorphic history of the region. The suggestion of a major Pliocene orogeny preceding peneplanation is entirely at variance with the sequence of events noted by other observers and summarized by Jutson (1934).

The faulting which, Teichert notes (1947*b*), breaks Permian and Cretaceous strata "into a number of westward-tilted blocks" must be of a date anterior to the peneplanation of regional extent, which Jutson regards as reaching perfection in the Miocene period. Upheaval occurred after this, probably in Pliocene times, developing broad undulation (also of regional extent) of a plateau surface; but this is broken in a few places by block faulting.

Woolnough (1920) regards the Stirling Range, in the south of Western Australia, as a major tectonic landform bounded by fault scarps 2,000 feet high; and block-faulted plateaus and sunklands separated and bounded by long and prominent scarps regarded as tectonic have been reported in the Northwest Division and in the Napier-Geikie Range in the far north.

The most remarkable scarp, however, which has been generally attributed with confidence to recent dislocation on a large scale is that of the Darling fault. This is a marginal feature of the land, not orogenic in the sense of bounding a horst or other form of block mountain uplift; but, if the topographic form has been produced by dislocation of the land surface, it is, nonetheless, a tectonic feature. The fault is roughly parallel to the west coast, toward which the scarp descends, at a minimum distance from it (near Perth) of 20 miles. For more than 200 miles, according to the account by Jutson (1934), this line is marked by a prominent scarp 800–1,000 feet high, of resistant ancient rocks, while at the base of this lies a so-called "coastal plain," consisting of thin blown sands and alluvia overlying Mesozoic strata.

Successive rejuvenations of rivers descending across the scarp have been described (Clarke and Williams, 1926). These rejuvenations are attributed to intermittent upheaval of the hinterland as the Darling fault scarp has been progressively developed, and geomorphic evidence of this kind has led to a presumption that a tectonic theory of origin is correct. Geologists are, however, still searching for evidence of dislocation sufficiently recent to satisfy them that the landscape feature has not been merely exposed by fault-line differential erosion and accentuated by marine erosion. Laterite on the foreground surface has, it seems, not been detached by faulting from the "duricrust" of the inland plateau above it (Prider, 1948).

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LEACHING OF CARBONATES IN GLACIAL DRIFT AND LOESS AS A BASIS FOR AGE CORRELATION[1]

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ABSTRACT

The rate of leaching of carbonates in Pleistocene deposits is affected by several factors, each of which is variable. Therefore, it is considered that correlation of two deposits of less-than-stage value stratigraphically, based on comparable mean depth of leaching in the two, is not likely to be reliable.

THE PROBLEM

It has long been recognized by geologists as well as by soils scientists that one of the conspicuous chemical processes in the formation of soil from raw material—such as glacial drift and loess—rich in calcium carbonate, is the gradual solution of calcium carbonate by downward-percolating water and its reprecipitation (especially under subhumid and semiarid climates) at greater depth below the surface. Both the leaching and the reprecipitation ordinarily take place above the water table. The dual process is particularly evident in the drift and loess in the Central Lowland region, for there the Pleistocene deposits include a large quantity of limestone and dolomite ground up by the glacial mill, and the climate is sufficiently moist to promote the active circulation of subsurface water. East of the Lowland the drift is generally less rich in carbonates; west of it the climate is too dry to permit rapid leaching.

Movement of calcium carbonate is one of many processes involved in the formation of soil from parent material. The relative intensities of these processes are controlled, in turn, by a number of basic factors. These controls have been grouped in various ways. According to Thorp (1941) they are grouped thus:

1. Parent-material
2. Climate
3. Biologic activity (the successive organisms that operated on and in the soil during its formation)
4. Relief (topographic position)
5. Time

To this list might be added minor factors in the geologic environment such as position of water table.

Modern opinion assigns a greater importance to climate, relative to time, than was formerly thought probable, and some soils are now believed to record the influence of climatic fluctuation during their formation (Bryan and Albritton, 1943). These fluctuations have commonly involved changes in vegetation covers and in the positions of water tables. Probably most of them are phenomena of the glacial and interglacial ages.

The belief has been held by many geologists that the thickness of the leached zone increases, with the passage of time, to a maximum imposed by average depth of water penetration. In this relationship, it has been reasoned, there may be a means of determining the relative ages of two or more drift sheets. Specifically, within one climatic zone and in drifts of similar composition, the differences in thickness of the leached zone in two different sheets of drift should indicate the relative ages of the two drifts. This view stresses the importance of the time factor and gives little consideration to the other factors that influence soil making.

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The evaluation of depth-of-leaching measurements demands an account of the leaching process and of the attempts made to use depth of leaching as a guide to stratigraphic correlation.

THE PROCESS

As rain water soaks into the ground, it is far from saturated with calcium bicarbonate (the substance present in the dissolved state) and it contains carbon dioxide derived from the air and from the humus zone at the surface. Calcite is readily soluble in water charged with carbon dioxide, and solution proceeds from the upper surface downward, until a state of saturation is reached. At the level at which saturation occurs, calcium carbonate is precipitated as a powdery filling of soil interstices, as concretions, and in some places as thick, rather continuous beds. Secondary carbonate is thereby added to the primary carbonate already present at this level.

The factors that control precipitation of secondary carbonates are not well understood. However, the chief process involved appears to be evaporation. In subhumid regions much downward-percolating water evaporates before it reaches the water table. Evaporation seems to occur in two distinct ways: (1) by ordinary evaporation within permeable soil, aided under suitable conditions by capillarity and in the drier regions by desiccation cracks; and (2) by transpiration of soil moisture by plants, the roots of which, even in subhumid regions, may penetrate several feet below the surface.

VARIABLE FACTORS

If we apply the factors known to influence soil formation in general to the specific process of leaching of carbonates, we can list a number of specific variable factors.

Parent-material: texture and stratification.—The conspicuous textural differences most commonly met with are the differences between clay-rich till and sandy till, and between till and loess. In the Central Lowland the tills commonly are rich in clay and very compact; hence their permeability to percolating water is relatively small. In contrast, the loess sheets consist predominantly of silt and are less compact than the tills; hence their permeability is greater than that of till. It has been estimated that in northwestern Illinois, loess undergoes leaching at about twice the rate applicable to till (Leighton, 1923, p. 280). In a stratified section, in which a body of till is overlain by a layer of surficial loess thinner than the leached zone, the thickness of the loess strongly influences the depth of leaching. In post-Iowan drift in southeastern South Dakota this situation occurs commonly, with 2–4 feet of loess overlying clay-rich till. Where surface drainage is poor, the loess is almost invariably leached from top to base; and the depth of leaching, if any, in the underlying till increases with decreasing thickness of the loess. In materials of comparable permeability, solution proceeds more rapidly in small particles of a given carbonate substance than in large particles.

Parent-material: composition—especially its content of primary carbonate.—It is evident that, other factors being equal, increase in thickness of the leached zone will take place at a slower rate in material rich in primary carbonates than in material poor in these compounds. The richer the material, the greater the total amount of solvent action that must occur per unit of thickness; in rich material the percolating water becomes saturated with calcium bicarbonate quickly; this reduces the rate of increase in the thickness of the leached zone (though it does not necessarily reduce the rate of the leaching process itself). The solution of dolomite, of course, takes place more slowly than that of calcite because of its lesser solubility.

Climate.—Significant climatic changes are known to have affected wide regions during the glacial and interglacial ages, including the time since the climax of the latest glacial age. Such changes may be expected to have influenced the rate of leaching directly through changes in precipitation and temperature and indirectly through changes in vegetation and in the position of the water table. Of these components, probably the most important is precipitation.

Other factors being equal, large precipitation, distributed more or less uniformly throughout the year, should result in a greater rate of increase in thickness of the leached zone than should small precipitation sharply concentrated in time. It is difficult to find comparative data because areas in the Central Lowland with markedly different precipitation are normally so far apart that the influence of other factors on depth of leaching cannot be minimized. Another complicating factor is the capillary movement of water upward, resulting in evaporation and in deposition of carbonates in the soils of areas having a mean annual precipitation of less than about 28 inches of water and with a well-marked dry season each year (Howard, 1946, 1947).

Character of vegetation.—It has been learned empirically that in some regions at least, and with other factors equal, thickness of the leached zone is greater beneath a forest cover than beneath a grass cover. This confirms the theoretical consideration that downward-percolating subsurface water beneath a forest is more acid than that beneath grassland. For example, in three counties in central Illinois, loess samples taken from beneath forest vegetation show a mean depth of leaching more than 50 per cent greater than that beneath grass vegetation (Smith, 1942, pp. 170–171).

Topography—especially surface slope.—It has been noted repeatedly that with other factors equal, the thickness of the leached zone becomes thinner with increase in the surface slope directly above. In fact, beneath thoroughly dissected topography possessing no nondrained or poorly drained areas, till and loess commonly exhibit no leached zone at all. This is true even in old drifts such as the Kansan. Presumably this condition is the result of the large proportion of local precipitation that runs off the surface. The large proportion of runoff not only reduces percolation into the soil and consequent solvent action but also permits mechanical erosion of the soil to take place, thereby reducing the thickness of the leached zone or inhibiting it altogether. In some places deflation also may have played a part in removing an unknown upper portion of the soil.

Position of water table.—Since leaching generally occurs only above the water table,[2] the water table ordinarily sets a lower limit to leaching. Where the water table is very close to the surface, leaching is not apparent. On the other hand, in subhumid regions the water table generally lies far below the zone of precipitation of secondary carbonates, and there is no apparent connection between the two features. From these statements it appears that in some situations the position of the water table may influence the depth to which leaching penetrates.

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[2] It is possible for leaching to occur below the water table if the ground water is not saturated with calcium bicarbonate.

Time.—It is not known with certainty whether increasing depth of leaching is a straight-line function of time or whether its relation to time is represented by a curve. However, we may deduce that the rate of increase of depth of leaching probably diminishes as time progresses, for two reasons. The first is the necessity of redissolving ever increasing amounts of secondary carbonate precipitated earlier. The second is the loss of downward-moving water abstracted through the roots of plants and transpired into the atmosphere, as the water passes through an ever increasing thickness of soil already leached. Whether or not this deduction is valid, the quantitative relation between depth of leaching and time is not known; hence the results of measurement of the leached zone yield values that at best are only relative.

SUMMARY OF VARIABLE FACTORS

It was stated near the beginning of this paper that attempts had been made to derive relative time figures from measurements of depth of leaching. However, consideration of the variable factors involved makes it clear that the results of such calculations must include a wide margin of error.

MINIMIZATION OF VARIABLES

Realizing the presence of variable factors, investigators have attempted to reduce them to a practical minimum. In consequence they have achieved results which, though not accurate, indicate distinct age differences between pairs of drift sheets known independently to have conspicuously different ages. Reduction of variable factors consists mainly of (1) selecting drift sheets of comparable composition and general texture, (2) selecting drift sheets in the same district with respect to rainfall, (3) selecting test points only on broad, flat uplands with negligible surface slope, (4) avoiding the comparison of data obtained from beneath a forest cover with those obtained from beneath a grass cover, and (5) using auger borings rather than surface exposures for sampling, so as to minimize the possibility of contamination. Of these points 1 and 3 are believed to be the most important and 5 seems the least important. The writer's samplings in the Wisconsin drift in South Dakota suggest that there, at least, it is almost negligible.

NECESSITY FOR MANY SAMPLINGS

Despite the exercise of care in reducing the variability of contributing factors, the resulting figures, within any district, show a wide range of variation among individual test points. Hence statistical means derived from many individual tests have had to be resorted to in order to show consistent differences between two drift sheets. In Indiana (Thornbury, 1940) 51 samplings of the thickness of the leached zone in Illinoian drift close to the border of Kay and Leighton's Tazewell drift showed a mean of 158 inches, but with an extreme range of 126 inches (102–228 inches) or about 80 per cent of the mean. Possibly this great range and that in the Tazewell drift cited below are in part the result of leaching under a variable thickness of loess. Corresponding figures for 64 samplings of the Tazewell drift, close to the same border, are: mean = 58 inches, extreme range = 54 inches (36–90 inches), or about 93 per cent of the mean. Although there is a gap of only 12 inches between the minimum figure for the Illinoian and the maximum figure for the Tazewell, the frequency distribution of the samplings[3] supports the validity of the mean figures given.

On the basis of the foregoing figures, thickness of the leached zone would seem to be a useful factor in distinguishing between stratigraphic units, such as the Illinoian and the Wisconsin drifts, having the value of stages. In particular, depth of leaching should be of value in fixing the position of the border of the younger drift in detail where, as is often the case, no end moraine is present at the border. At any rate, probably it would be of value only within a geographically restricted region.

The mean value of the Illinoian drift samplings in Indiana (158 inches) contrasts with a mean value of 150 samplings of the same drift sheet in western Illinois, of only 97 inches (Leighton, 1923). The most obvious variable involved here is mean annual precipitation, which is about 40 inches in the Indiana district sampled and only about 34 inches in the Illinois district sampled. Whether or not the marked difference in depth of leaching is to be correlated with the difference in precipitation or with other factors such as thickness of overlying loess, the variability of results from district to district is noteworthy as a barrier to the useful employment of leaching-depth data in the determination of age differences.

Within a single drift sheet, the Tazewell, large groups of samplings were made between pairs of successive end moraines presumably distinguished by minor age differences. In Indiana the results (Thornbury, 1940, p. 454) were as shown in table 1. The frequency distribution of the individual samplings cannot be assessed because only the means are given. The extreme range of these *means* is 20 inches (38–58 inches), a sizable proportion of the mean thickness derived from the four figures given in table 1. The data suggest, as was pointed out by the investigator who obtained them, the reasonable inference of gradual decrease in the age of the Tazewell drift from south to north. But they suggest also that, to be useful in correlation, such data could be drawn only from a drift sheet so thoroughly studied that the individual minor layers of which it is composed, corresponding to individual end moraines, are known in detail.

TABLE 1

Areal and Stratigraphic Position	Leached Zone*
Immediately behind the Shelbyville moraine (mean of 64 samplings) . . .	58
Shelbyville drift (mean of 181 samplings)	51
Champaign drift (mean of 111 samplings)	42
Bloomington drift (mean of 100 samplings)	38

* Mean thickness in inches.

WISCONSIN DRIFT IN ILLINOIS AND IOWA

The results obtained in northwestern Illinois (Leighton, 1923) and eastern Iowa (Leighton, 1923; Kay and Graham, 1943, pp. 107–108, 195–196; Alden and Leighton, 1917, pp. 81–82[4]) illustrate the difficulty encountered in attempting to correlate units having the small value of substages as contrasted with stages. The data are summarized in table 2. The Iowan and Tazewell substages are fairly well separated from each other on a basis of mean thickness. On the other hand, the drift of the Green River lobe, on the basis of these figures alone, would certainly be classed as Iowan. Yet Leighton (1923, p. 280) found the loess overlying the till of the Green River lobe to be notably thicker than the loess overlying either the Iowan or the Tazewell drifts. This difference, largely the result of the geographic position of the Green River lobe, influenced the leaching figures significantly because loess, being more permeable than till, permits more rapid leaching than does till. Taking this factor into account, Leighton judged the correlation of the drift of the Green River lobe to be indecisive; later he (Flint and others, 1945) dated it as Tazewell.

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[3] Individual samplings are not given in the original reference; W. D. Thornbury kindly supplied two sets of unpublished individual values.

[4] The data from Alden and Leighton are not comparable with the data from the other sources because they are derived from leaching depths measured not from the surface but from the top of the till beneath a capping of loess.

TABLE 2

Substage and Areal Position	Leached Zone*
Iowan drift in eastern Iowa:	
146 samplings (Leighton)	62
37 samplings (Kay and Graham) . .	66
Tazewell drift in northwestern Illinois:	
Belvidere lobe, 120 samplings	48
Bloomington drift, 56 samplings . . .	44
Drift of undetermined age:	
Green River lobe, 119 samplings . . .	68

* Mean thickness in inches.

In an excellent report on a detailed examination of Tazewell drift in northern Illinois, Willman and Payne (1942, pp. 176–177) presented a table of data on the thickness of the leached zone but avoided correlation of these data with specific drift sheets, apparently because of the variable factors involved. The table clearly shows that depth of leaching is greater where surficial silt (presumably loess) overlies sand or gravel than where it overlies till, because silt is much more permeable than till.

Samplings of two drifts of post-Iowan age in southeastern South Dakota made by the writer in 1946 and 1947 yielded the results given below. The two drifts are of no more than substage value. The older drift is covered with 2–5 feet of loess, whereas the younger is not; otherwise the factors seem comparable.

Twenty-five samplings of the older of the two drifts (Cary?) near its border in Lincoln, Minnehaha, Turner, and Bon Homme counties gave a mean depth of leaching of 29 inches with an extreme range of 22 inches (18–40 inches). Fifty-eight samplings of the younger of the two drifts (Mankato?) near its border in Hutchinson, Yankton, and Bon Homme counties gave a mean value of 23 inches, with an extreme range of 48 inches (0–48 inches). The frequency-distribution plots made from the two sets of data fail to show a close grouping near the mean; in consequence it seems doubtful that the arithmetic means are reliable indicators of age difference between the two drifts.

Eighty-four samplings of the older drift (Cary?) in Charles Mix County, just west of the counties named above, gave a mean of 15 inches with an extreme range of 36 inches (0–36 inches). The frequency-distribution curve is very irregular and unsatisfactory; hence the mean figure is believed to have little value. The irregularity is thought to be the result of variable composition of the till, which contains Niobrara chalk sharply varying in amount from place to place. The drifts sampled in the counties to the east contain relatively little chalk and are more nearly uniform in composition. In this case, therefore, variable lithology is the chief disadvantageous factor.

CONCLUSION

The data given from South Dakota, added to the more extensive data cited from Indiana, Illinois, and Iowa (which include nearly all the detailed published references), are believed to support the conclusion deduced in the body of this paper, namely, that variable factors preclude the accurate determination, through depth of leaching, of the relative time intervals that have elapsed since the deposition of any two drifts. The data show, however, that in at least some pairs of glacial-drift sheets, time differences corresponding in stratigraphic value to stages—as the difference between the Illinoian stage and the Wisconsin stage—although inaccurate, are reflected in depth-of-leaching figures. For time differences of lesser value the inaccuracies created by variable factors make for results of doubtful reliability.

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GEOLOGICAL NOTES

LIQUID INCLUSIONS IN GRANITE THERMOMETRY[1]

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INTRODUCTION

Microscopic liquid inclusions have been used by a number of investigators to indicate temperatures of crystallization. This method of thermometry consists of expanding the liquid by heat until the contraction vacuole normally present in each inclusion disappears. The temperature at which the vacuole disappears is the original temperature of inclusion, neglecting pressure considerations.

The method assumes (1) that the liquid completely filled its cavity at the time of crystallization; (2) that the liquid is an aqueous solution uniform in concentration from inclusion to inclusion within a single crystal and containing no carbon dioxide; (3) that the pressure on the liquid at the time of inclusion was small or that its magnitude can be estimated; (4) that primary liquid inclusions can be distinguished from secondary inclusions; (5) that there has been no significant change in the volume of the cavity itself due to pressure, solution, or precipitation; (6) that there has been no addition or loss of liquid from the cavity; and (7) that the samples used are representative of the deposit under study.

These assumptions have been discussed in recent works of Newhouse (1933), Ingerson (1947), and Twenhofel (1947). The assumptions appear to be justified in certain cases where moderate temperatures and pressures are involved. This paper reports an attempt to extend the method to the quartz of several selected granites, in the formation of which higher temperatures and pressures may be involved.

APPARATUS

The essential apparatus for investigating liquid inclusions consists of a heating stage for the microscope, a heating medium, and a temperature indicator. Air was found to be the most satisfactory heating medium for studying high-temperature inclusions. Its only disadvantage is that the rock section must be highly polished to see into the interior. In the apparatus employed, a slow-moving stream of dehydrated air under constant pressure passes through an alundum tube containing a nichrome wire-resistance coil. The heated air then passes through a thin, insulated cell of sheet tin, which rests on the microscope stage. The cell has an inside diameter of 0.6 inch and an inside depth of 0.2 inch. The section to be studied rests on two horizontal rungs in the center of the cell, so that the air flows equally over the top and the bottom. Transmitted light enters the cell through a glass plate in the base. A thick cover glass fits over an opening in the top. When necessary, air jets set in the side of the insulating container may be played on the microscope objective to keep it cool.

During operation, each rock section is heated gradually to the temperature of vacuole disappearance. The controlled-current flow in the heating coil regulates the temperature of the air flow. Once equilibrium has been established, the temperature of the air flow remains constant for a given amperage. Ten minutes are allowed for the rock section to reach the temperature of the air stream.

Temperatures are measured by a thermocouple, which may be calibrated with re-

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spect to steam and to the melting-points of standard tin, bismuth, lead, and zinc metals. Temperature readings are taken by inserting the thermocouple through a small hole in the side of the cell and placing the tip in contact with that portion of the section being studied. Readings of both the base and the top of the section are made, and the average of the two readings is taken as the temperature of the section. With a section of optimum thickness the difference is not over a degree or two at 400° C.

It has been found advantageous to use a fairly thick rock section rather than a standard 0.03-mm. thin section. Besides providing more inclusions in one section, the thicker plate allows better determination of the relationships of individual inclusions to planes and of planes to cracks or crystallographic surfaces. On the other hand, a thick section is less transparent and requires longer to reach equilibrium with the air current. The optimum thickness for work with quartz was found to be about 0.2–0.4 mm.

A magnification of three to four hundred times is required for work with the small inclusions in granites. A fairly large working distance is recommended, however, in order to avoid overheating the microscope objective. The writer used a 21-power objective and a 20-power hyperplane ocular.

DISCUSSION OF RESULTS

Sections suitable for liquid-inclusion analysis were prepared from fifty-two different granites from the University of Wisconsin petrographic collection. Nine of these granites were selected for detailed work because of the abundance of inclusions, the clarity of the quartz grains, or a favorable geologic history. For example, the Wausau (Wis.) granite was selected because of the extreme abundance of liquid inclusions. The Barre (Vt.) granite has few inclusions but is probably free of later metamorphism. The Mellen (Wis.) sample came from an inch-wide granite veinlet in gabbro. The veinlet must have been liquid and cannot be regarded as a product of granitization.

The liquid inclusions present are smaller in diameter than those in most hydrothermal or pegmatite quartz bodies. Under low magnification they appear as cloudy layers and speckled bands. Under higher magnification these specks resolve into individual inclusions which lie in or close to definite planes and curving surfaces.

Differentiation between primary- and secondary-type inclusions proved extremely difficult. Primary inclusions are those which form at the same time as the enclosing crystal. They are normally considered to be related to growth or crystallographic planes and directions; but this relationship may not be evident unless the crystallographic planes and directions are marked by zonal lines, color bands, solid inclusions, or other evidence. Because quartz is the last constituent of granite to form, it fills the available interstices and seldom shows crystal form or zonal growth lines. Secondary inclusions are those introduced subsequent to consolidation of the portion of the crystal in which they occur. Laemmlein (1929) has shown that they may form by the sealing-off of fractures containing a liquid. Dale (1923) suggests that the rift and grain of granites are due to closely spaced, strain-induced sheets of secondary liquid inclusions in the quartz. It is also possible that secondary inclusions are due to later re-working or recrystallization.

No true criteria exist for differentiating primary and secondary inclusions in the absence of crystal outlines and zonal growth lines. Planes of primary inclusions never cross one another and do not cross grain boundaries. Planes of secondary inclusions, however, may cross one another and primary planes as well. They may or may not cross grain boundaries. Planes of secondary inclusions may be superimposed upon planes of primary inclusions. Similar orientation of planes of inclusions over an entire section and gradation into fractures point to a secondary origin. Inclusions with branching, tubular shapes are secondary; but inclusions with other shapes, including those of negative crystals, are indeterminate. Optical orientation is not readily applicable to planes in quartz, yielding, at best, only their positions relative to traces of the prism faces, which are parallel to the *c*-axis. Planes of inclusions, therefore, can be related to crystallographic directions but not to definite crystallographic planes.

By use of these criteria, at least half the liquid inclusions examined could be classed as definitely secondary. The remainder could not be classified. None were definitely primary.

TABLE 1*

RESULTS OF LIQUID-INCLUSION THERMOMETRY: TEMPERATURE OF VACUOLE DISAPPEARANCE

Locality								
Wausau, Wisconsin	Mellen, Wisconsin	Barre, Vermont	Knorre, Meissen, Saxony	Ane, Saxony	Rhyolite Mtn., Cripple Creek, Colorado	Grouse Mtn., Cripple Creek, Colorado	Petershead, Scotland	Rockport, Mass.
170° C. (1) 240 (1) 300 (1) 226 (2) 243 (1) 267 (2) 216 (2) 225 (1) 260 (3) 270 (2) 340 (1) 200 (4) 220 (4) 240 (2) 260 (1) 216 (2) 225 (2) 248 (1) 172 (1) 196 (1) 330 (1)	222° C. (1) 248 (1) 260 (2) 305 (1) 235 (1) 276 (1) 298 (1) 350 (1) 198 (1) 230 (2) 258 (1) 289 (2) 310 (1) 320 (1) 360 (1)	208° C. (1) 216 (1) 232 (2) 298 (1) 300 (1) 338 (2) 340 (1) 368 (1) 384 (1)	192° C. (1) 200 (2) 208 (2) 215 (1) 241 (1) 260 (2) 310 (1) 223 (1) 251 (1) 280 (2) 200 (1) 206 (1) 215 (2) 223 (1) 225 (2) 240 (3) 260 (1) 270 (2) 291 (1) 300 (1) 320 (1) 225 (1) 230 (1) 256 (1) 260 (1) 267 (2) 280 (2) 300 (1) 317 (2)	195° C. (1) 215 (2) 221 (1) 250 (2) 278 (1) 280 (1) 292 (1) 330 (1) 358 (1) 227 (1) 243 (1) 260 (2) 285 (1) 200 (1) 242 (1) 245 (1) 276 (1) 304 (1) 310 (1)	205° C. (2) 220 (1) 235 (1) 218 (1) 235 (1) 250 (2) 278 (2) 280 (1) 335 (1) 180 (1) 200 (1) 230 (2) 280 (1)	192° C. (1) 231 (1) 252 (1) 160 (1) 200 (1) 220 (1) 223 (1) 250 (1) 298 (1)	245° C. (2) 262 (1) 283 (1) 286 (1) 330 (1) 195 (1) 218 (1) 265 (1) 278 (1) 240 (1) 245 (1) 260 (1) 275 (1)	216° C. (1) 233 (1) 250 (2) 228 (1) 241 (1) 262 (1) 265 (1)

* Inclusions are grouped according to the planes in which they occur. Numbers in parentheses refer to the number of inclusions.

Because no primary forms could be distinguished, a number of both indeterminate and secondary-type inclusions were heated until the vacuoles just disappeared. The results are listed in table 1, in which the inclusions are grouped according to the planes in which they occur. Temperatures are accurate within the range ±3° C. No pressure corrections have been made.

The resulting figures are all relatively low, but they are primarily important for showing that there is considerable variation in the temperatures at which the vacuoles disappear, even within the same plane of inclusions. In all the fifty-two granites examined, no planes could be found whose inclusions showed any visual uniformity in the size of vacuole relative to cavity. No zoning in temperature within or between planes was noted; instead, vacuoles in adjacent inclusions differed by as much as 163° C. in their temperature of disappearance. No carbon dioxide content was noted in any inclusion.

Many inclusions containing huge vacuoles, indicating leakage of liquid, were not run. Their temperatures would be considerably higher than any of those listed. Leakage during heating was also common. Inclusions showing leakage were detected and eliminated by measuring the size of every vacuole at room temperature both before and after application of heat to the section.

No conclusions as to temperatures of crystallization can be drawn from table 1. The irregularity of observed temperatures within individual planes of inclusions indicates that at least one of the assumptions of the method is not justified in the present case. Variations in temperature may be due to leakage, to inclusion of a gas phase either above or below the critical temperature, or to effects of the little-understood processes which cause secondary inclusions. Perhaps primary liquid inclusions are rare or lacking in granites. If they are present, better criteria are needed for their recognition. Secondary inclusions may form within the range of primary crystallization temperatures, but this cannot be proved until secondary and primary inclusions can be distinguished.

The present investigation emphasizes the difficulties attendant upon the method of liquid-inclusion thermometry in granites. The method may yet succeed, but exceptional care will be required in selection of specimens, in evaluation of inclusions, and in technique. It is also possible that the prevalent concept of the crystallization process of quartz must be modified. If crystallizing igneous quartz passes through a desiccating gel stage, irregular inclusions might be expected. Liquid inclusions forming within a viscous mass may shrink or swell irregularly, especially under conditions of nonuniform stress or of resurgent solutions.

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SOURCE AREA OF GREAT PLAINS PLEISTOCENE VOLCANIC ASH[1]

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INTRODUCTION

Known occurrences of the Pearlette volcanic ash zone of Pleistocene age (early Yarmouthian) are scattered from west-central Texas to southeastern South Dakota and from western Kansas as far east as south-central Iowa (Frye *et al.*, 1948). The volcano and silicic lava field associated with this immense outburst of acid glass must have been of giant size. The prevailing wind direction indicates that it may have been situated somewhere to the southwest of the area of Pearlette ash. The Capulin group of volcanoes has been suggested as a possible source (Landes, 1928), but the cones are comparatively small, and the lavas of the Capulin region consist of basalts, andesites, trachytes, soda trachytes, and phonolites (Collins and Stobbe, 1942). However, these volcanoes are closer to the areas of Pearlette ash than are any other Pleistocene volcanoes, and hypothetical acid tuff and pumice from the Capulin group may conceivably

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have been completely buried under later flows. Other possibilities which should not be arbitrarily dismissed are the large areas of extrusive rocks in northwestern United States, although the prevailing winds are not from the northwest. A more favorable source is the Valle Grande volcanic area of north-central New Mexico. The distances from this area to the known localities of Pearlette ash range from 220 miles (Hartley County, Texas) to 740 miles (Minnehaha County, South Dakota). Their locations are shown in figure 1.

VALLE GRANDE

The Valle Grande region lies in Sandoval County about 30 miles west of Santa Fe and is part of the Jemez Mountains, between the Rio Grande on the east and the Nacimiento Mountains on the west. It is shown on the

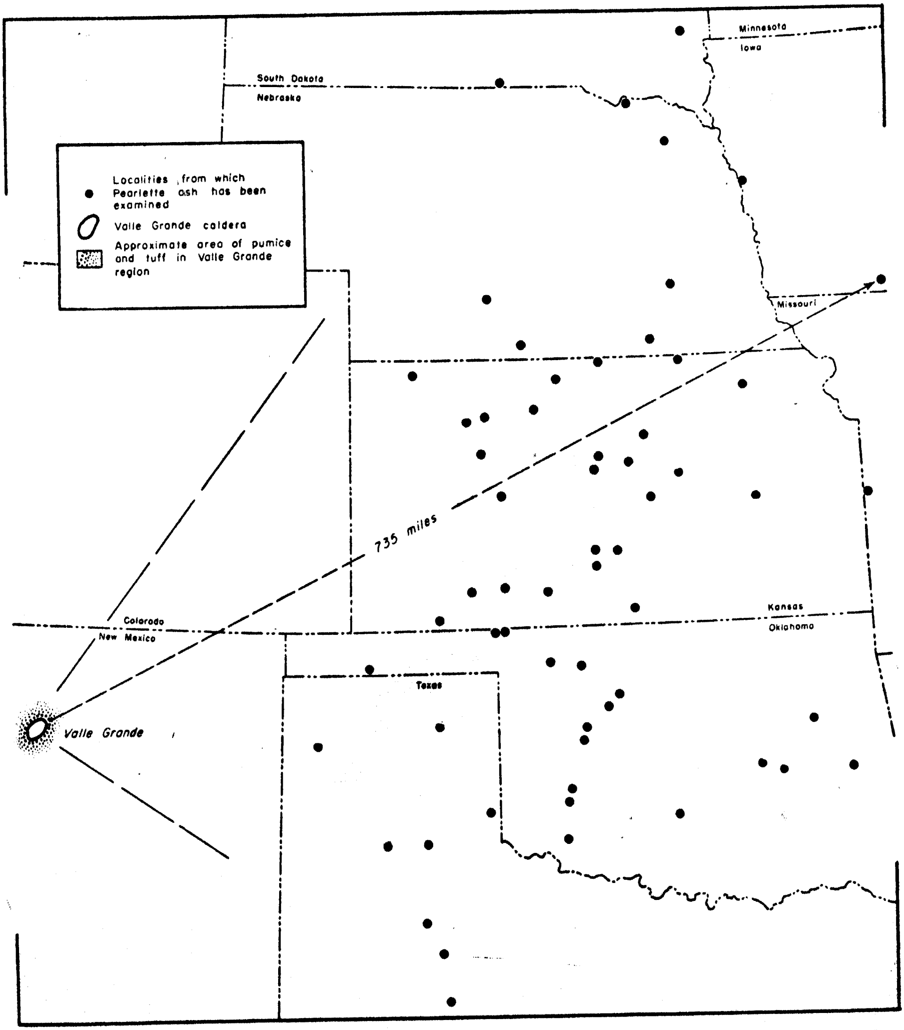

FIG. 1.—Map showing Valle Grande crater and localities from which Pearlette volcanic ash has been examined. Not more than one ash locality is plotted from each county.

Jemez, Jemez Springs, and Santa Clara topographic sheets and is illustrated on the Raisz (1939) landform map as the Sierra de los Valles. The feature is described by Ross (1938) as follows: "This is an explosive vent or rather group of vents, that have built up a crater of elliptical shape, about 13 by 17 miles in extent. This makes it the largest crater ever reported. The materials ejected are almost exclusively tuffs which reach a thickness of at least 1,000 feet on the east flank." Williams (1941, pp. 251–252) regards the volcano as a caldera of collapse origin.

The rim is surrounded by an area about 30 miles in diameter, consisting of rhyolite, tuff, and pumice (Talmage and Wootton, 1937, p. 41). The tuff beds on the eastern flank form the Pajarito Plateau, which is cut by radiating canyons that reach a depth of 1,000 feet in tuffs (Ross, 1931). At least one of these canyons, Frijole, in Bandelier National Monument, is the site of ancient cliff dwellings, and the tuff of this canyon and its topographic expression have been described by Henderson (1913). On the west side of the caldera is a related broad sloping plateau of tuff and pumice which extends to the east flank of the Sierra Nacimiento (Renick, 1931, p. 70). Workable pumice deposits are also found as far north as Rio Arriba County (Clippinger, 1946, pl. 1). A generalized geologic map of the area has been compiled by Church and Hack (1939, p. 617).

The tuffs, which overlie the late Tertiary Santa Fe formation, include a dozen distinct beds (Williams, 1941). They are judged by Ross (1931) to be comparatively recent in age and perhaps as late as early Pleistocene. Earlier workers (Lindgren *et al.*, 1910, p. 151) assigned a Miocene age to the material, and, still earlier, Reagan (1903, p. 104) considered it to be pre-Cretaceous.

PETROGRAPHY OF THE TUFF AND PUMICE

The pumice and tuff are cream colored to light gray. The tuff forms vertical walls of canyons and weathers to a hard buff or brown surface. Iddings (1890, pp. 10–12) first described the petrography of the tuffs, from samples collected by J. W. Powell in 1887. He also described two varieties of rhyolite from the area: a normal porphyritic variety and a mica-bearing rhyolite; and he concluded that the tuff was composed of the former rhyolite. He described the tuff as carrying porphyritic crystals of quartz and sanidine and small fragmentary crystals of augite. In some of the tuffs the quartzes are rounded, and in others they are dihexahedral pyramids with short prism faces. Many of the sanidines exhibit a blue iridescence (Iddings, 1890, p. 10).

Iddings describes pumice related to the mica-bearing rhyolite, which includes quartz and sanidine phenocrysts and also oligoclase. The mica consists of small flakes of biotite, and the quartz is in crackled grains and has an amethyst color.

For the present study tuff was collected from the cliff on the north side of Frijole Canyon near the Tyuonyi ruin (samples 1 and 2) and from a road cut between the museum and the checking station in Bandelier National Monument (sample 3). Pumice was collected from a road cut on State Highway 4 on the northeast side of the Valle Grande caldera (sample 4) and from the same highway on the southeast side of the caldera (sample 5).

Petrographic examination of the glassy part of the tuff and pumice was made according to methods previously described (Swineford and Frye, 1946). The glass of tuff sample 1 has the same pale-buff color (Ridgway 17‴f) and refractive index (1.500 ± 0.001) as the Pearlette. The shapes of the shards are similar to Pearlette shard shapes, but sample 1 contains many more fibrous shards. It also has phenocrysts, up to 1 or 2 mm. in diameter, of subangular to rounded quartz and sanidine. There are a few crystals of other minerals, particularly magnetite; no biotite was observed.

Sample 2 is somewhat weathered, so that the original color is destroyed. The shards are similar in shape to those of sample 1, but the refractive index of the glass is 1.502 ± 0.001. There are also amber-colored euhedral quartz phenocrysts of the habit described by Iddings and somewhat weathered sanidine phenocrysts. The sample contains a small quantity of tiny magnetite grains.

Sample 3 is an almost neutral gray (Ridgway 15''''f), has an index of about 1.52, and contains many large quartz and sanidine phenocrysts. Much of the sanidine shows the blue iridescence or sheen noted by Iddings, and many of the quartz crystals have sharp interfacial angles. Crystals of other minerals are rare.

Pumice sample 4 is fibrous and frothy in appearance, and its color and index are the same as those of the Pearlette. It contains small phenocrysts of quartz and sanidine. In this sample, mineral grains were separated by the use of bromoform diluted with alcohol to a specific gravity of about 2.45 in order to remove the quartz and feldspar as well as the heavier minerals from the glass. Heavy minerals include magnetite, hornblende, augite or diopside, hypersthene or enstatite, pale-green olivine, apatite, a trace of corroded biotite, zircon, and others. The crystals are enclosed in glass and constitute approximately 8 per cent of the sample, by weight.

TABLE 1

CHEMICAL COMPOSITION OF PUMICE AND TUFF FROM VALLE GRANDE REGION, AND PLEISTOCENE VOLCANIC ASH FROM THE GREAT PLAINS

(Analyses by Russell Runnels)

No.	Location	SiO_2	Al_2O_3	Fe_2O_3	CaO	MgO	TiO_2	SO_3	P_2O_5	Ignition Loss	Diff.	Moisture 105°–140°
Valle Grande:												
1	Frijole Canyon	66.77	15.68	2.18	0.71	0.65	0.43	Nil	.....	6.17	7.41	0.30
3	Road cut, E. side of caldera	74.81	13.01	1.79	1.36	1.56	0.12	Tr	Tr	0.42	6.93	.....
4*a*	NE. side of caldera	70.45	14.25	1.86	1.92	0.90	0.63	Tr	0.08	3.08	6.83	0.99
4*b**	NE. side of caldera	71.32	14.02	1.36	1.67	0.82	0.37	Tr	Nil	2.90	7.54	0.85
5	SE. side of caldera	70.50	13.66	1.96	1.74	0.82	0.71	Tr	Tr	0.42	10.19	.....
Pearlette:												
1†	Minnehaha Co., S.D.	71.41	14.00	1.73	0.78	0.06	.....	.....	.....	4.12	7.90	0.30
3†	Harrison Co., Ia.	70.98	13.01	1.96	1.16	0.22	.....	.....	.....	3.91	8.76	0.29
7†	Seward Co., Neb.	71.29	13.29	1.81	0.82	0.21	.....	.....	.....	4.06	8.52	0.30
20†	Gove Co., Kans.	71.68	13.20	1.93	0.93	0.13	.....	.....	.....	4.21	7.92	0.21
42†	Custer Co., Okla.	72.53	14.14	1.34	1.19	0.37	.....	.....	.....	4.20	6.23	.....
48†	Briscoe Co., Tex.	70.80	12.40	1.69	0.56	0.32	.....	.....	.....	4.16	10.07	0.09

* After removal of crystalline fraction.

† Analyses from Frye *et al.* (1948).

Pumice sample 5 is a pale cream color (Ridgway 17'''g), somewhat lighter than typical Pearlette. Its refractive index is 1.502 ± 0.001, and the glass is similar to sample 4 in general appearance. It contains large anhedral grains of clear colorless quartz and sanidine, pseudo-hexagonal biotite books, dark greenish-brown hornblende, and many small particles of magnetite, all of which can be recognized under the binocular microscope.

Of the five samples examined, 1 and 4 are judged to be possible correlatives of the Pearlette. As the phenocrysts and most of the heavy minerals should settle out from the air within a few miles of the volcanic vent, the properties of only the glass need be considered. Although the tuff (1) and pumice (4) contain more fibrous material than does typical Pearlette, the proportion of fibrous shards may also be a function of selective transport. The brown color of the weathered and case-hardened tuff has no counterpart in Pearlette deposits, but it may be attributed to the presence and alteration of hornblende, magnetite, and other iron-bearing minerals in the crystalline part of the tuff.

For further comparison with the Pearlette, chemical analyses of some of the Valle Grande samples were made. These are summarized in table 1, in which are also included some typical analyses of Pearlette ash. The pumice sample 4 was analyzed as a whole (4*a*) and also after removal of the crystalline constituents (4*b*). The latter analysis should be comparable with those of the Pearlette. Except for the slightly larger percentages of calcium and magnesium oxides in the New Mexico pumice, there seems to be no significant difference in composition between the Valle Grande material (sample 4) and the Pearlette ash. Sample 1, which is petrographically similar to the Pearlette, is somewhat low in silica.

CONCLUSIONS

Too few samples were collected at Valle Grande to determine the number of different tuffs, their stratigraphic order, and their relative volumetric importance. Geologists who are more familiar with the region are better qualified to answer these questions. Valle Grande is judged to be a probable source of the Pearlette volcanic ash for the following reasons: (1) Only an extremely large caldera, emitting highly acid lavas, such as Valle Grande, could supply ash which is as extensive as the Pearlette. (2) The volcano is situated to the southwest of the known Pearlette lentils. (3) The petrographic character and chemical composition of the glassy part of some of the tuff and pumice are similar to those of the Pearlette. (4) The tuffs are said to be early Pleistocene in age.

It is predicted that detailed work in the area may show a predominance of the Pearlette type of pumice and may definitely link it to the Pearlette ash zone of the central Great Plains and glaciated region.

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DISCUSSION

CRUSTAL MOVEMENT IN THE GREAT LAKES AREA: A DISCUSSION[1]

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Under the above title a paper has appeared in *Bulletin of the Geological Society of America* by Sherman Moore, engineer of the United States Lake Survey Office at Detroit (1948; see also Moore, 1922). He has examined the records of lake levels at all the gauging stations on the Great Lakes, which records cover, on the average, a period of fifty-two years, though some extend over a considerably longer time. This is much the most complete study of its kind for the area, and it confirms all the earlier studies from Gilbert to Gutenberg in showing that the Great Lakes area has been tilted so that its northern (really east-northeastern) side is higher than the southern one, and that this tilt is still continuing (Gilbert, 1898; Freeman, 1926; Gutenberg, 1933).

Moore has, however, offered the startling and wholly unwarranted interpretation that this canting of the earth's surface is due to a subsidence of the area—a downtilting toward the south, not an uptilting toward the north. This is contrary to all that has been learned by Gilbert (1898), Goldthwait (1907), Hobbs (1911), Leverett and Taylor (1915), Bergquist (1936), Stanley (1936, 1937), and other glacialists during the last half-century, from measurements (precise levels) of the now abandoned beaches of the Maumee, Whittlesey, Warren, Algonquin, and Nipissing glacial lakes—all ancestors of the present Great Lakes and formerly in occupation of the area.

So far as the gauge records alone are concerned, a downtilt to the south or an uptilt to the north would produce identical changes in the water levels measured at the gauging stations. This can be illustrated by holding a half-filled basin of water in a horizontal position, raising one side through a certain distance, returning it to a horizontal position, and then lowering the opposite side by the same amount. The movement of the water surface with reference to the sides of the basin will be exactly the same, whether the basin is uptilted, as in the first instance, or downtilted, as in the second.

It will help to make clear that the crustal movement has been an uptilt if reference is made to the map of figure 1, which shows the outlines of glacial Lake Nipissing, the latest of the glacial lakes, whose outline was little different from those of the present Great Lakes, except that Lake Superior was then a part of the Michigan-Huron basin. In time Lake Nipissing followed several earlier glacial lakes, the latest of them Lake Algonquin, during the existence of which the hinge line of no vertical movement had been migrating northward. The present elevations of the isobases of the Nipissing shore line, measured in feet above mean tide, are shown on the map by the figures enclosed in parentheses, and the actual amounts of uplift at each of them during the last two thousand or more years are shown by the figures not so enclosed. It will be noted that, in the Lake Superior area, Moore's hinge line of no vertical movement and his isobase of maximum subsidence very nearly correspond in position with the isobase of maximum uplift and that of the hinge line of no uplift, respectively.

The rise of the water level at Duluth, which Moore has interpreted as due to a subsidence, is explained not by subsidence but by *greater* uplift of 100 feet at the north shore of the lake. This has caused the water to be withdrawn from that side and has raised the level at the southwestern shore. A similar withdrawal of water from the northeastern shore of Lake Erie, where the outlet is likewise to the northeast, and a submergence of the southwestern shore have taken place for the same reason (Moseley, 1905; Hobbs, 1911, pp. 27–35). No such changes have occurred in Lake Huron because its outlet is at the south end of the lake, so that the displaced water passes out through the outlet.

The withdrawal of the water from the rock-cliffed north shore of Lake Superior and the

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[1] Manuscript received August 10, 1948.

submergence of its southern shore are strikingly displayed to the boat traveler along them. On the north shore the bottom of the lake is laid bare for some distance out from the now abandoned cliffed shore (Lawson, 1893). On the southern shore in all places to the south of the isobase, which passes through the outlet at Sault Ste Marie, the shore is submerged (Stuntz, 1870; Wright, 1905; Hobbs, 1911, pp. 27–35).

If we consider the beaches of the earlier glacial Lake Algonquin, they show a far greater amount of uplift. Over the same area, the rate of their average uplift is 2.6–3 feet per mile, as against 0.5 foot for the Nipissing Lake, which we have been considering (Stanley, 1939).

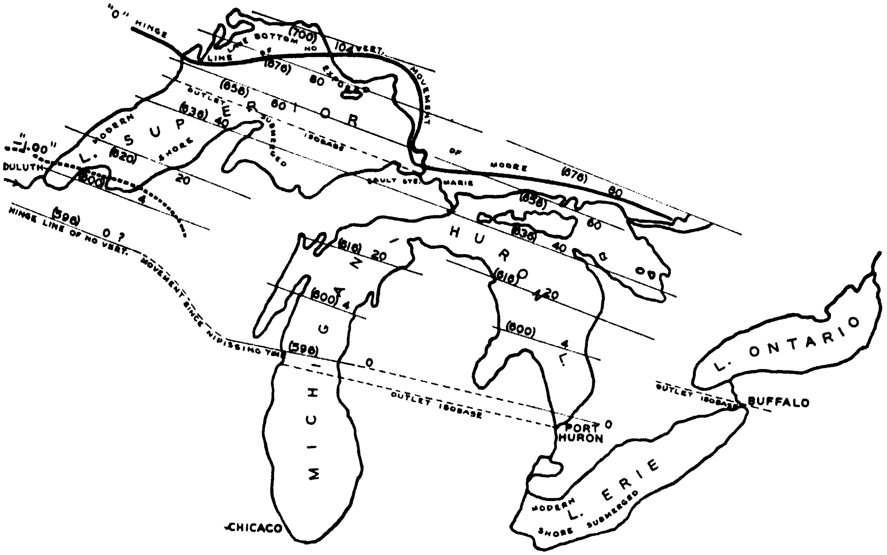

FIG. 1.—Map showing isobases of the now warped Nipissing water plane and the amounts of actual uplift. The hinge line of Moore and his isobar of maximum depression are shown by the heavy and dotted sinuous lines (map mainly after Taylor).

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ANCIENT ARCTICA: REMARKS AND SUPPLEMENT[1]

WOLF MAYNC
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As A. J. Eardley's paper "Ancient Arctica" (1948) is incomplete and in part in error with respect to eastern Greenland, some corrections and additional information are offered.

My detailed stratigraphical studies in 1936–1938 of the post-Devonian sediments in eastern Greenland justify a note on some errors and omissions in Eardley's paper. Although recent works on the stratigraphy of Alaska, the Arctic Archipelago, Spitzbergen, etc., are referred to in Eardley's compilation, the geologic literature on Greenland from 1939 onward is, unfortunately, not taken into consideration. In his bibliography Eardley lists papers presented at the East Greenland Meeting in Schaffhausen, Switzerland, on March 11–12, 1939; but the data contained in these contributions have not been incorporated in his article.

In order to give an idea of what has been achieved during the past years with regard to the geology of eastern Greenland, a list of recent papers is given at the end of the present note.

This list of references is supposed to stimulate interest in the geography of eastern Greenland, which, beyond all doubt, may today be considered the best-known part of the Arctic regions.

Although the so-called "Carboniferous" deposits of eastern Greenland and of other Arctic regions have been shown to be Middle to Upper Permian in age (Aldinger, 1935; Maync, 1938, 1939, 1940, 1942), they were still designated "Carboniferous" by Eardley. These beds, the distribution of which is, moreover, wrongly shown on the paleogeographic map (Eardley, 1948, fig. 7, p. 427), were proved to be in part Zechstein formations, lithologically and biofacially identical with the "Magnesian limestone" and "Marl slate" of Great Britain, the German Zechstein deposits, and the Russian Kungurian-Kazanian sediments of Pae-Choi, Kanin Peninsula, Timan, Pinega Basin, Kazan, etc. (Maync, 1938, 1939, 1940, 1942). Furthermore, these reefal Zechstein deposits of eastern Greenland were found to be stratigraphically associated with fossiliferous *Productus* limestone and ammonoid-bearing *Martinia* limestone (with *Cyclolobus, Medlicottia*) of undoubted Upper Permian age (Maync, 1938, 1939, 1940, 1942; Miller and Furnish, 1940). Another facies member of the eastern Greenland Permian is the *Posidonomya* shale. These shales carry a ganoid fish fauna which contains diagnostic Zechstein forms known from the German *Kupferschiefer*, the English "Marl slate," and the Kazanian of Russia (Aldinger, 1935, 1937). This fish fauna, moreover, shows close affinities to that of the Middle Phosphoria of Wyoming, a formation which is proved by its cephalopod faunas to be equivalent to the Word (and Capitan?) formations of Texas (Middle to Upper Permian), and by its foraminiferal fauna to be of Zechstein age.

The occurrence of these genuine Zechstein deposits in eastern Greenland, of course, proves a marine connection with England and Germany; and the wide distribution of Permian (formerly Carboniferous or Permo-Carboniferous) beds all around the Arctic region suggests that the present-day North Polar Sea was the center of this late Permian Arctic Basin. A broad gulf (Scandic Sea) extended southward between Greenland and Spitzbergen across the British Isles and terminated in the inland Zechstein Sea of Germany (Koch, 1935; Maync, 1940, 1942). The "Scandinavian-Scottish Continent" (Stamp, 1947; Eardley, 1948) of the Lower Carboniferous had ceased to exist. Its late Paleozoic remnant was the Scandinavian-Baltic Shield, which apparently has never borne any marine Permo-Carboniferous deposits. We

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[1] Manuscript received November 6, 1948.

cannot share Eardley's view, therefore, that "the southern extent of the Carboniferous basin on the east coast of Greenland is a matter of speculation" (1948, p. 427): Nowhere in eastern Greenland did any Carboniferous sea encroach upon the continental border, and even the supposed marine "Upper Carboniferous" beds on Holms- and Amdrups Land (80°–81° N. lat.) are younger than Lower Permian (Artinskian) and correspond more or less to the Kungurian stage of Russia. The eastern Greenland beds are deposits of the Upper Permian sea, which is traceable south into the heart of Europe. Another arm, branching from the Permian Arctic Basin, ended in the Russian Kungurian-Kazanian seas. The discovery of *Cyclolobus*, the diagnostic Upper Permian ammonoid genus of the Himalayan geosyncline in the eastern Greenland Permian obviously discloses that marine connections existed between the Arctic Basin and the far-eastern geosyncline (Himalayas, Ussuri Bay, Mongolia, Siberia).

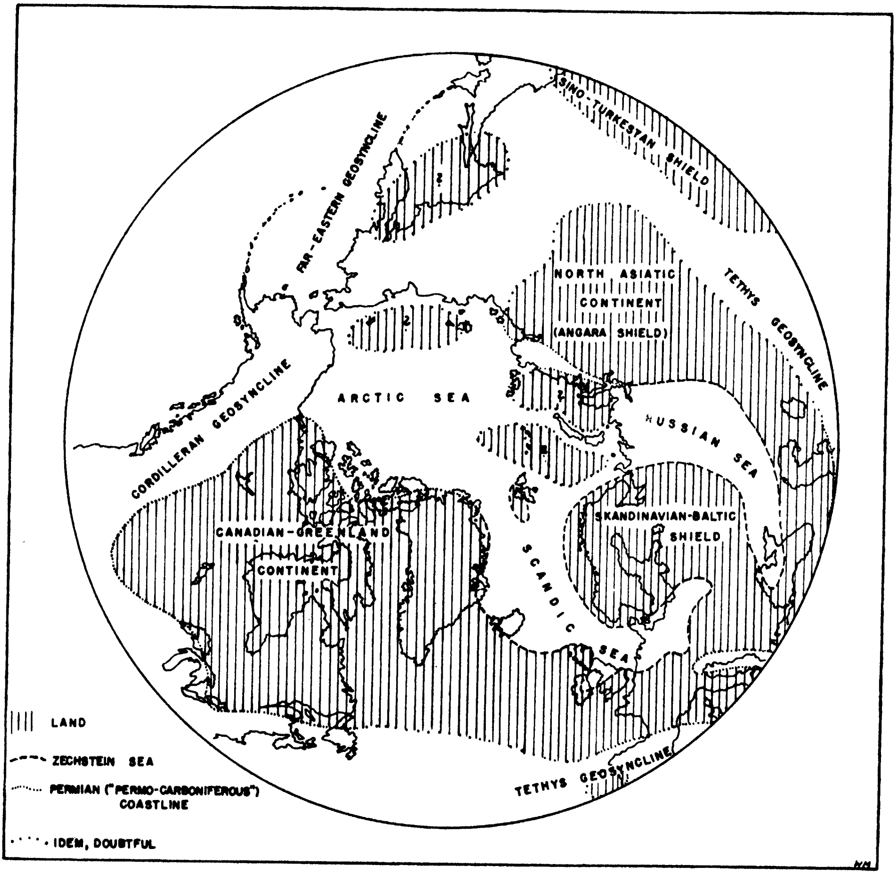

FIG. 1.—Paleogeographical map of the Arctic regions in Permian ("Permo-Carboniferous" or "Carboniferous" *auctorum*) times (after Eardley, 1948; Gignoux, 1943; and Maync, 1940, 1942).

Figure 1 depicts the supposed distribution of land and sea during the Permian ("Carboniferous" or Permo-Carboniferous *auctorum*) as a supplement to Eardley's paleogeographical map (1948, fig. 7, p. 427).

Information on the Triassic formations of eastern Greenland is furnished by the comprehensive papers of A. Rosenkrantz, L. Koch, E.

Nielsen (see Koch, 1935), L. F. Spath (1935*a*), and H. Stauber (1942).

The Jurassic faunas of southeastern Greenland have been dealt with by L. F. Spath in a masterly manner (Koch, 1935; Spath, 1935*b*, 1936), and a detailed paper on the Jurassic stratigraphy of eastern Greenland and the paleogeographical relations with other Arctic regions was published recently (Maync, 1947). A publication by L. F. Spath (1947) is devoted to Jurassic and Cretaceous ammonite faunas from southeastern Greenland. Until now, beds of Bathonian-Callovian, Neo-Oxfordian, lower Kimmeridgian, upper Kimmeridgian, and Volgian (Portlandian) age are known to occur in northeastern Greenland. Facial and faunal evidence shows that the Scandic Sea also existed in the Jurassic, and the landmass between Greenland, Spitzbergen, and Scandinavia postulated by Eardley seems to me, therefore, improbable or at least very disputable.

Erosion after the late Cimmerian block-faulting has removed part of the eastern Greenland Jurassic, and synorogenetic boulder beds of the Volgian locally overlie different portions of the older Jurassic series with a sharp angular unconformity.

A monograph dealing with the Cretaceous stratigraphy of eastern Greenland and the Arctic regions was finished some years ago, but its publication was considerably delayed because of postwar difficulties in Denmark (Maync, 1948*d*). It is to be hoped that its contents may furnish the Greenland data needed for a paleogeographical map of the Arctic region. The above-mentioned paper contains paleogeographical maps of northeastern Greenland in both Valanginian and Aptian-Albian (and Senonian) times and attempts to correlate the circumarctic Cretaceous deposits.

In northeastern Greenland the occurrence of fossil-bearing sediments of Infravalanginian-Rjasanian (Subcraspeditan), middle Valanginian (Polyptychitan), Aptian-Albian, and lower Senonian age has been ascertained (Koch, 1935; Frebold, 1935; Maync, 1938, 1939, 1940, 1948*c*, *d;* Stauber, 1938, 1939, 1940). Farther south, beds of Cenomanian-Turonian age have recently been recorded (Spath, 1946).

While the Lower Cretaceous faunas of eastern Greenland still bear a distinct boreal character, proving the existence of a large isolated Arctic *Aucella* sea, the Aptian-Albian faunas contain cosmopolitan forms which are found in Cretaceous beds all over the world. This fact shows the establishment of marine connections between the Arctic sea, the Mediterranean Tethys, and the Pacific geosynclines.

Eardley's figure 8 (p. 429, lower map), showing the distribution of the Arctic coal-bearing "Miocene" (*recte* Eocene) deposits, does not contain the eastern Greenland type locality on Sabine Island, discovered in 1869–1870 by the German Koldewey Expedition, the flora of which was dealt with by O. Heer in his famous *Flora fossilis arctica* (1868–1883). The stratigraphical work in 1936–1938 disclosed several new localities with Tertiary plant- or coal-bearing beds on Sabine and Pendulum islands, Wollaston Foreland, and Hold-with-Hope (Maync, 1938, 1939, 1940). These Tertiary sediments were deposited during the interval between the first basaltic extrusions and the overlying basalt flows (traps).

Another famous locality of early Tertiary plant-bearing beds, viz., Atanikerdluk on Nugssuak Peninsula, western Greenland, is sought for in vain on Eardley's map.

Since the publication on L. Koch's paper (1935), referred to by Eardley in discussing the tectonics of eastern Greenland (p. 431), several important works have been published which have thrown new light on the structure of eastern Greenland and its history (see the papers by Bierther, Bütler, Cloos, Schaub, Stauber, Vischer, Wegmann, listed at the end of this article).

Koch's idea of strong Tertiary faulting along the eastern coast of Greenland is not strictly correct because all the major faults recorded are Paleozoic and Mesozoic in age and have only in part been later reactivated. Since the Variscan orogeny, no folding has taken place in eastern Greenland, but crustal tensions led to an intense block-faulting of the continental border.

The main phase of this step-faulting falls within upper Kimmeridgian times: the different tectonic blocks faulted en échelon (*antithetische Schollentreppe*) underwent a westward tilting, so that tectonic grooves were formed in the joints of the tilted blocks (see fig. 2). These depressions and the adjacent fault scarps have played an important role during the late Jurassic and Cretaceous sedimentation (marine troughs and basins bordered to the west by steep, unstable fault cliffs). Furthermore, intra-Cretaceous movements of some of these old blocks are recorded (post-Valanginian, post-Albian, pre-Tertiary).

The few critical remarks given in the forego-

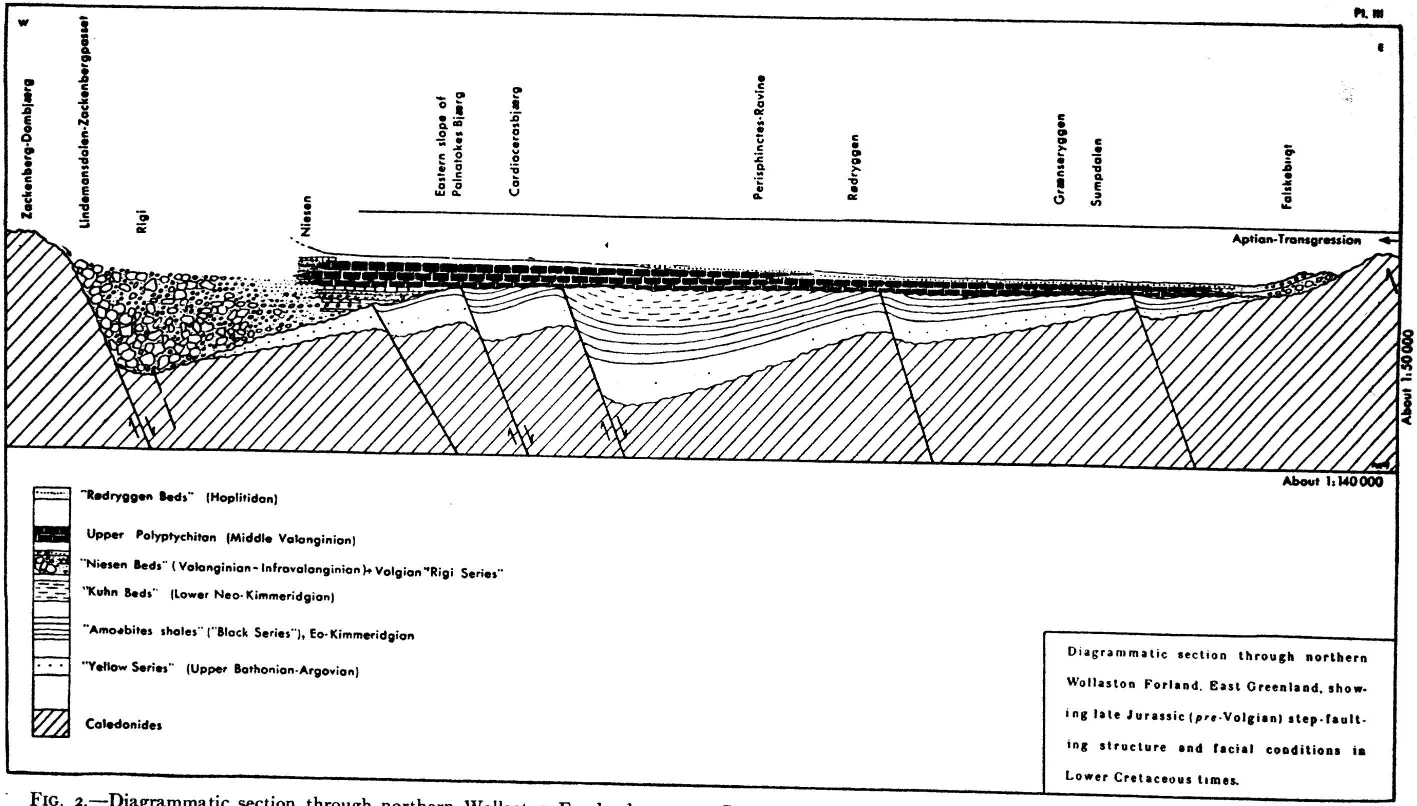
Pl. III
W
E
Zackenberg-Dombjærg
Lindemansdalen-Zackenbergpasset
Rigi
Niesen
Eastern slope of Palnatokes Bjærg
Cardiocerasbjærg
Perisphinctes-Ravine
Rødryggen
Grænseryggen
Sumpdalen
Falskebugt
Aptian-Transgression
About 1:50 000
About 1:140 000
"Rødryggen Beds" (Hoplitidan)
Upper Polyptychiton (Middle Valanginian)
"Niesen Beds" (Valanginian-Infravalanginian)+ Volgian "Rigi Series"
"Kuhn Beds" (Lower Neo-Kimmeridgian)
"Amoebites shales" ("Black Series"), Eo-Kimmeridgian
"Yellow Series" (Upper Bathonian-Argovian)
Caledonides
Diagrammatic section through northern Wollaston Forland, East Greenland, showing late Jurassic (*pre*-Volgian) step-faulting structure and facial conditions in Lower Cretaceous times.

FIG. 2.—Diagrammatic section through northern Wollaston Foreland, eastern Greenland, showing late Jurassic (pre-Volgian) step faulting structure and facial conditions in Lower Cretaceous times (Maync, 1948*b*, pl. 3).

ing are not intended to discredit the whole of Dr. Eardley's paper, the importance of which is duly acknowledged. I know from experience how difficult it is to embark upon so delicate an enterprise as the compilation of paleogeographical maps of little-known regions. I wish only to point out some omissions and misinterpretations of the region that I happen to have studied at first hand. The rapid advances in the knowledge of eastern Greenland's geological history during the last ten years certainly deserve careful consideration. The main shortcomings in Eardley's article are due chiefly to ignorance of the essential and indispensable literature.

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[2] Largely published after L. Koch's synopsis, *Geologie von Grönland* (1935).

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ANCIENT ARCTICA: REPLY TO REMARKS[1]

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It is unfortunate that I was not more thorough in searching the literature of East Greenland in my article on Ancient Arctica, and missed the publications that appeared during and after the war. It is evident that Mr. Maync's kind remarks should be taken as an authoritative supplement to the article and should be carefully integrated into it by those interested in Arctic paleogeography.

This is an opportune time to call attention to still another important and recent publication not referred to in the article or by Mr. Maync, namely, one by L. R. Wager, entitled "Stratigraphy and Tectonics of Knud Rasmussens Land and the Kangerdlugssuaq Region." It deals with the great plateau basalts and fossiliferous interbeds and the deformation they have suffered. The first marine transgression in the area occurred in Senonian time, and then immediately afterward volcanic activity broke out to form the "lower lavas and tuffs," which may in part be lowest Eocene. The greatest igneous event then followed and at least 20,000 feet of basalt was poured out to give the "plateau basalt series." It is all probably lower Eocene in age. Shallow-water sediments below and above show that sinking, compatible with the thickness of the basalts, occurred during or soon after their extravasation. Immediately afterward, the Plateau Basalts were cast into a grand flexure which is now along the coast with the ocean-

[1] Manuscript received January 4, 1949.

ward side down. The differential vertical movement between Greenland and the Greenland Sea basin immediately adjacent is 20,000–25,000 feet. Although the basalts are not present north and south of the middle east Greenland area, the topography along the coast, according to Wager, suggests comparable movements. This recent evidence of profound vertical movements in the Greenland area is strikingly in accord with the general theory of the article on Ancient Arctica, viz., that the past geographies of the Arctic and north Atlantic were vastly different than now, that the changes are due largely to vertical movements of the crust in the magnitude of 5,000–25,000 feet, and that the crust of the great region is one of continental composition—not true ocean floor, such as parts of the Pacific are. The new evidence is much more significant than the faults mentioned in the article, many of which, as Maync points out, are very ancient, and whose significance to the theory is doubtful.

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REVIEWS

Geological Map of the Province of Ontario: Map No. 1946–3. Compiled by J. SATTERLY. Ontario Dept. of Mines, 1946. 1:1,267,200 or 1 inch to 20 miles, 40 × 51 in. 50 cents.

Geologists will welcome the publication of a new geological map of one of the largest and most complex of the Canadian provinces. The extensive mapping program of the Ontario Department of Mines and other agencies during the last two decades has made compilation of this map a necessity. The scale of the map permits delineation of more detail than is shown on the map of North America. The map of Ontario, therefore, is a useful contribution to the geological literature of that province.

Dr. Satterly exercised commendable restraint in his compilation. No effort was made to make the Archean rocks of the different parts of the province fit into any restricted system of nomenclature or correlation. The Archean is divided into only volcanic rocks (principally Keewatin), sedimentary rocks (Timiskaming and other similar series), and basic (Haileyburian) and acidic intrusives.

The Proterozoic rocks are separated into three catagories, namely, sedimentary (Bruce and Cobalt), sedimentary and volcanic (known locally as Animikie, Whitewater, or Keweenawan), and basic intrusives (mainly Keweenawan). The post-Cobalt granitic intrusives (Killarney), unfortunately, are not separated from the Archean; and it is not clear even from the map legend that there is a post-Huronian granite. The term "Huronian" is missing, although the type area lies within the province.

Rocks of four Paleozoic systems (Cambrian through Devonian) are appropriately shown, as is also the Cretaceous.

A synoptic summary of the mineral production of the province to the end of 1945 is included on this sheet.

F. J. P.

Geological Map of Saskatchewan, Map 895A. Canada Department of Mines and Resources, Mines and Geology Branch, Bureau of Geology and Topography, 1947.

This map, on a scale of 1 inch to 20 miles, is another in a series of generalized provincial geological maps. The geology is based for the most part upon the work of the Geological Survey of Canada, with some information from Imperial Oil Limited. The Drafting and Reproducing Division is responsible for the cartography.

Somewhat more than 50 per cent of the map has been colored, but an additional 10 per cent is indicated to be underlain mainly by Upper Cretaceous formations. Rocks of pre-Cambrian age constitute the bedrock of about 40 per cent of the province, Mesozoic and Cenozoic strata about 57 per cent. Productive metal mines, as well as developed and undeveloped water-power sites, are indicated by appropriate type.

The legend includes 32 map units, of which 13 are pre-Cambrian, 3 Paleozoic, 12 Mesozoic, and 4 Cenozoic. The colors appear to be well chosen and, for the most part, match very well those of the Manitoba sheet (*Map 850A*). Several of the colors for pre-Cambrian rocks, however, registered somewhat lighter than did those of the corresponding divisions on *Map 850A*. The map units are indicated by number as well as by color, and it is here that difficulties arise. For example, map unit 19 in Saskatchewan is the Lower Cretaceous Clearwater formation, while on the adjoining Manitoba sheet the same number, but different color, is used for the Upper Cretaceous Vermilion River formation. It seems to this reviewer that a more desirable system would be one based on letters, similar to that of the United States Geological Survey.

Detailed work in the large region underlain by pre-Cambrian rocks has been confined chiefly to such areas of economic importance as Lake Athabaska, Lac La Rouge, and Amisk Lake. The difficulties of correlation are apparent in a glance at the legend: map units 1 and 3 are both chiefly volcanic rocks, 2 and 4 are chiefly sedimentary rocks, while 5 represents volcanic and sedimentary rocks. The terms "Keewatin" and "Timiskaming" are not used in the legend. Archean or Proterozoic age is suggested for the succeeding three map units, respectively: 6, complex of sedimentary, volcanic, and intrusive rocks; 7, chiefly basic intrusive rocks; and 8, chiefly acidic intrusive rocks. It is in this group that some discrepancy with the Manitoba sheet was noted. Along the Churchill River at the Manitoba boundary, *Map 895A* records chiefly sedimentary rocks associated with acidic intrusions (*6c*—orange). The adjacent portion of the Manitoba sheet shows undifferentiated granitic intrusions and older gneissic and schistose sedimentary and volcanic rocks (*7c*—pink). It seems to the reviewer that the Saskatchewan map should have been colored pink (*8a*) rather than that the Manitoba map should be colored orange. This is the only place where the two maps do not match in color. With the exception of map unit 13, which designates late Proterozoic diabase masses of limited extent, rocks of definite Proterozoic age are confined to the Lake Athabaska—Cree Lake region. Alcock (1936) has mapped in detail the area north of Lake Athabaska, recognizing the Beaverlodge series of sediments (9), basic intrusives (10), and a Proterozoic granite (11). The Athabaska series (12), comprising coarse clastics with minor basalt, constitutes about 10 per cent of the entire map sheet and among the pre-Cambrian rocks is second only to the acidic intrusives in areal extent.

Paleozoic rocks are represented by relatively small areas of color, but the boundaries of the possible extensions of the map units are indicated by an inconspicuous dotted line. The three map units represent generalized groupings as follows: 14, Upper Ordovician; 15, Silurian; and 16, Devonian. No formational names are given.

Mesozoic rocks are distributed among twelve map units, nearly all of Upper Cretaceous age. Map unit 17, found as small scattered patches, represents the Swan River group of Lower Cretaceous and possibly earlier age. Map units 18, the McMurray formation, and 19, the Clearwater formation, both of Lower Cretaceous age, are shown only along the Clearwater River near the Alberta boundary. The Ashville formation, unit 20, listed as "Lower (?) and Upper Cretaceous," is found only near the Manitoba boundary. The remaining eight map units are all labeled "Upper Cretaceous" in age. They include the following formations in ascending sequence: 21, Favel formation; 22, Vermilion River formation; 23, Lea Park formation; 24, Ribstone Creek, Grizzly Bear, and Birch Lake formations; 25, Oldman formation, Pale and Variegated beds; 27, Bearpaw formation; and 28, Eastend, Whitemud, Battle, and Frenchman formations. Map unit 26, the Riding Mountain formation, is designated as correlative with part of the Lea Park, together with the succeeding formations up to and including the Bearpaw. It is recognized in the eastern half of the province, the more complicated nonmarine and marine succession being found in the western half. It is to be regretted that, in listing the formations in both units 24 and 28, the legend carries the succession in inverse order, the youngest beds being placed at the bottom. It is of interest to note that the name "Alberta formation" does not appear in the legend, although it is widely used in the literature and has been recognized by Wickenden (1941) as correlative with the Favel and part of the Vermilion River formations.

Strata of Cenozoic age are mapped in four units, only two of which are at all extensive. The first of these, 29, the Ravenscrag formation, is mapped as Paleocene in the Cypress Hills region, where it has been redefined by Furnival (1946, p. 106). Elsewhere the same color is used with the symbol *29a*, where some Upper Cretaceous beds are included. The Swift Current beds, 30, of Eocene age occur only as isolated patches southeast of the city of the same name. The Cypress Hills formation, 31, of Oligocene age is the second map unit of any extent. It is confined to the southwest quarter of the sheet. The symbols 32 and 32*A* are used for the Wood Mountain formation of Miocene age, consisting chiefly of gravels. These represent the youngest rocks in the province, other than the unconsolidated deposits of Pleistocene age.

This map and the companion sheet, *Map 896A*, comprise a set which has already been found useful for instructional purposes by the reviewer. It is expected that it will be widely used. The cartographers have been able to include considerable detail without sacrificing clarity and legibility, and they have produced an excellent work. Additional maps of this type will be most welcome.

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H. S. Armstrong

Mineral Map of Saskatchewan, Map 896A. Canada Department of Mines and Resources, Mines and Geology Branch, Bureau of Geology and Topography, 1947.

This map is the companion sheet, on the same base, of the geological map of Saskatchewan (*Map 895A*). The eight mining districts into which the province is divided are outlined. Metallic mineral properties are indicated by a large spot and clearly legible number, both in red. The numbers are distributed through an alphabetical index of fifty mining properties, corrected to July, 1946. Metallic mineral occurrences are designated by a small red × with the name of the mineral in small red capitals. Nonmetallic mineral occurrences are marked in the same way, but in blue print. Economic deposits of clay of the Cretaceous Whitemud formation are shown by solid blue patches. In the pre-Cambrian portion of the map numerous glacial striae are shown by an appropriate arrow symbol.

This sheet will be found to be a useful complement of the geological map. It should be of particular interest to students of both economic geology and geography.

H. S. Armstrong

Geological Map of Manitoba, Map 850A. Canada Department of Mines and Resources, Mines and Geology Branch, Bureau of Geology and Topography, published 1946.

This map, on a scale of 1 inch to 20 miles, is one of several provincial maps which have been issued in recent years. The geology is compiled from the work of the Geological Survey, the Manitoba Department of Mines and Natural Resources, and the Ontario Department of Mines. The cartography is by the Drafting and Reproducing Division.

Most of the area between latitudes 49° N. and 55° N. has been colored; but, with the exception of the area west of Granville Lake, the remainder of the province northward to its boundary (60° N.) is colored only along some of the major waterways. The names of Robert Bell, J. B. Tyrrell, D. B. Dowling, and William McInnes are associated with this mapping, most of it done prior to the turn of the century. Productive metal mines and deposits of nonmetallics are indicated by distinctive symbols and type.

The legend includes twenty-two map units, of which eight are pre-Cambrian, six Paleozoic, five Mesozoic, one Cenozoic, and two of indefinite age. Most of the colors chosen to represent the different units are easily recognizable. If the geological structure were more complex, however, difficulty might be found in distinguishing the somewhat similar shades of green used for rocks of such diverse ages as Archean, Ordovician, and Cretaceous.

Rocks of pre-Cambrian age, comprising the larger part of the map sheet, are divided into eight units. The first three, of Archean age, are as follows: (1) chiefly volcanic rocks, (2) chiefly sedimentary rocks, and (3) volcanic and sedimentary rocks. This threefold division of the Archean rocks, a lithologic rather than a stratigraphic grouping, reflects the lack of sufficient detail in some of the source material. It is noteworthy that the terms "Keewatin" and "Timiskaming" do not appear in the legend, although Lawson's original "Keewatin" area lies just east of the southeast corner. Possible Archean or Proterozoic age is suggested for each of the succeeding four map units: (4) chiefly sedimentary rocks, (5) complex of sedimentary, volcanic, and intrusive rocks, (6) chiefly basic intrusive rocks, and (7) chiefly acidic intrusive rocks. The alternative ages suggested recognize the disagreement existing among Canadian geologists concerning the age of the sediments of unit 4. The intrusives of unit 7 are further subdivided and include gray sodic types as well as red potassic types. This subdivision is reminiscent of the similar twofold division of granites in northern Ontario (Moore and Charlewood, 1930). The only rocks to which a Proterozoic age is definitely assigned are the sediments, mostly quartzites, of the Churchill formation, colored as map unit 8.

Palaeozoic strata form a broad central band extending northwesterly from the International Boundary. Rocks of this age occur also along several rivers west and southwest of Hudson Bay. The Ordovician, Silurian, and Devonian systems are represented; but in only one of these —the Ordovician—is any further subdivision made. The earliest Paleozoic rocks are Upper Ordovician, comprising, in ascending order, the following formations: Winnipeg (9), Red River (10), and Stony Mountain (11). These three formations are mapped separately only to the south of latitude 52° N.; northward the map shows Upper Ordovician undivided (12). The Stonewall formation (13) forms a well-defined band immediately west of the Ordovician rocks.

The Middle Silurian age of these dolomites might well have been indicated in the legend. Devonian rocks (14) are adjacent to the Silurian and are suggested to be mainly Middle Devonian, with possibly some Upper Devonian beds. No mention is made of the formational names Elm Point, Winnipegosan, and Manitoban which have been applied to these strata (Kindle, 1912).

The age of the Amaranth formation (15), occurring just west of the Devonian, is not definitely known. It is younger than Devonian but older than Jurassic. A recent discussion of the correlation of this formation is given by Wickenden (1945).

Strata of undoubted Mesozoic age, mostly Cretaceous, are mapped in the southwest portion of the province. Map unit 16 represents Lower Cretaceous and earlier beds, including the Swan River group of Lower Cretaceous age. The later Cretaceous rocks form the Manitoba escarpment and its dip-slope. The Ashville formation (17), of Lower (?) and Upper Cretaceous age, is succeeded by the Favel formation (18), the Vermilion River formation (19), and the Riding Mountain formation (20), all of which are Upper Cretaceous in age. The Riding Mountain formation constitutes by far the largest area of Mesozoic rock in the map sheet.

The Boissevain formation (21), of Cretaceous or Tertiary age, and the Turtle Mountain formation (22), of Paleocene age, underlie a small area on the International Boundary at longitude 100° W. The Turtle Mountain formation represents the youngest consolidated rock in the province.

The map, which maintains the high standard of the Drafting and Reproducing Division, is an important addition to the series of more generalized Canadian geological maps. As such it should prove to be of considerable interest in the classroom and laboratory and will probably circulate widely.

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H. S. ARMSTRONG

The Ancient Volcanoes of Oregon. By HOWEL WILLIAMS. Eugene: Oregon State System of Higher Education, January, 1948. Pp. x+55; figs. 9; pls. 13. $1.25.

This recent work by Professor Williams, of the University of California, presents in printed form a series of lectures delivered under the auspices of the Oregon State Board of Higher Education. The lectureship was established in 1944 in honor of Dr. Thomas Condon (1822–1907), the first professor of geology at the University of Oregon.

Although written in popular style for the general reader, *The Ancient Volcanoes of Oregon* contains a concise summary of Williams' views on volcanism and therefore is of interest to geologists. The work also presents an excellent summary of the volcanic history of Oregon since the Cretaceous period. In the introduction the genesis of magma within volcanic regions is discussed. The magma below volcanic vents of the central type is arrested in its upward ascent within a few miles of the earth's surface and forms a reservoir by "displacing the surrounding rocks." The magma reservoir then becomes the magma-feeding chamber of the volcano. If eruptions of a volcano are frequent and the magma is continually supplied from the deeper-seated basaltic layer of the earth's crust, the products of the eruption will be dominantly basaltic. If, however, the magma-feeding chamber is cut off from replenishment by basaltic magma from below, it may and frequently does begin to crystallize prior to extrusion, with the consequent formation of a gas-rich, highly siliceous liquid phase, according to the crystallization-differentiation theory of N. L. Bowen. The composition of the extruded products will then be determined both by the extent to which crystallization differentiation has progressed and also by the depth or level at which the magma-feeding chamber is tapped by fissures extending to the surface. Among the minor causes of the diversity of lavas, Williams recognizes solution or assimilation of the wall rock of the magma-feeding chamber.

Williams discusses the types of volcanic eruptions and correlates these with forms of volcanoes. The introductory section is concluded with a brief résumé of the growth of Parícutin Volcano, Mexico.

The main section considers the volcanic history of Oregon from the Late Cretaceous to the present. In the discussion of the Recent epoch, which in Oregon covers approximately the last twenty-five thousand years, the sequences of events for Oregon's principal volcanoes are summarized. These include Mount Mazama (Crater Lake), The Three Sisters, Newberry Volcano, Diamond Craters, and others. In the discussion of The Three Sisters' region, no mention is made of E. T. Hodges' ancient supervolcano, Mount Multnomah (University of Oregon, 1925).

Although noting eruptions of Mount Shasta in 1786, Mount St. Helens between 1831 and 1854, Cinder Cone near Lassen Peak in 1851, and Lassen Peak itself from 1914 to 1917, Williams believes that the eruptions of the Recent epoch represent a marked decline in the volcanic history of Oregon and that the volcanic cycle is drawing to a close. The recent domes of rhyolitic obsidian representing the end-product of magmatic differentiation are suggested as evidence for an old-age decadent stage. Also the deeply eroded land forms common in the Cascade Mountains of Oregon suggest that forces of degradation are greatly in excess of those of construction. Yet some further outbursts from historically active volcanoes of the Cascade Mountains and possibly others can be expected in the future.

Williams' belief in waning volcanic activity in the Cascade Mountains may be regarded with some skepticism by critical readers. It should be remembered that, even though the accumulation of volcanic products within the twenty-five thousand years since the end of the Pleistocene epoch has been relatively minor, there were several million years in the Pliocene and Pleistocene for the volcanic pile which forms the Cascade Mountains to accumulate. Possibly more than one volcanic cycle may be recorded within this pile, judging from the diversity of volcanic products and the many erosional unconformities contained therein. If this is so, the commencement of a new volcanic cycle at any one of several Cascade volcanoes remains a distinct possibility, though probably a remote one in terms of a human lifetime.

F. M. Byers, Jr.

Gesteine und Minerallagerstätten. Vol. 1: *Allgemeine Lahre von den Gesteinen und Minerallagerstätten.* By P. Niggli. (Unter besonderer Mitarbeit für Teil IV von E. Niggli.) Basel: Verlag Birkhäuser, 1948. Pp. 540; figs. 335. Sw. fr. 46.

This book, in itself a store of information massed into 540 pages of solid-set type, represents only the first volume of a trilogy on rocks and mineral deposits. Volume 2, "Die exogenen Gesteine und Minerallagerstätten," is said to be in manuscript, and volume 3, "Die endogenen Gesteine und Minerallagerstätten," is in preparation.

The first volume is hardly a suitable textbook for students; it is more of a handbook of fundamental facts essential to the understanding of the sciences of physical geology, petrology, and mineralogy. It is divided into five parts: (I) the substance matter (*die stoffliche Grundlage*): geochemistry, crystal chemistry, and structural mineralogy; (II) the fabric: structural and textural properties of rocks (petrofabrics) and their relation to the technical properties; (III) the physical chemistry of the formation of mineral assemblages: primarily the thermodynamics of silicates and silicate melts; (IV) geophysical data pertinent to the understanding of rocks and mineral deposits; temperature and pressure conditions in the crust of the earth and mechanical, electrical, and magnetic properties; (V) the elements of classification and systematology of rocks and mineral deposits. Although the works of other investigators are frequently mentioned, no references to the literature are given. This is a definite shortcoming, and there is little consolation in the promise that volumes 2 and 3 will contain adequate references.

Never before has a book of this scope been attempted, and no geologist other than Niggli could have completed successfully a task so diverse and manifold. His superior insight and intimate knowledge not only of the geological sciences but also of chemistry, physics, and mathematics make the book an excellent source of information.

Although lack of space makes it impossible for the reviewer to discuss many details, it should be mentioned—for the benefit of those who know, or think they know, Niggli's stand on the question of granitization—that Niggli now willingly accepts this process. He says, for example (reviewer's translation):

> In addition to the metamorphism without material transport we have the metamorphism with more

> pronounced exchange of matter. The magmatic melts in the depth are able to emit matter which locally or regionally will penetrate the rocks of the lithosphere. . . . Thus truly mixed rocks will form . . . [p. 501]. Rocks which look very much like granite may likewise form through ultra-metamorphism, or even through ordinary contact metamorphism (e.g., from an arkose). If the differences are negligible, or if no change of name is justified, then one may speak of *metagranites* (granitic rocks formed through ordinary metamorphism), or of *ultrametagranites* (especially migmagranites or diabrochogranites). For better definition, and in order to avoid confusion, it may be expedient, when various bodies are present in the same region, not to use the simple word "granite" but, more precisely, to refer to the magmatic rock as *magmagranite* [p. 560].

In so far as the reviewer can judge, this unique book gives highly needed, correct, exact, and thoroughly reliable information, and it is, therefore, highly recommended to every physical geologist. However, if the American reader wants to profit fully, he should have some advance knowledge of the scientific jargon used by the Niggli school. There is a true jugglery with the symbols *c*, *fm*, *t*, *alk′*, *alk*, etc.; and a simple thing like the chemical composition of the feldspars is explained as follows (p. 72):

> Es sind Alumosilikate mit ähnlich gitterhaftem Tetraedergerüst wie die SiO_2-Modifikationen, jedoch teilweisem Ersatz des Si durch Al. Allgemeine Formel $[B^{IV} O_{4/2}]_n A_m$. In B^{IV} ist das Verhältnis Si : Al = 3 : 1, wenn A = K oder Na ist und 1 : 1, wenn A als Ca (eventuel Ba, Sr) auftritt. Auf 1 Al sind zur Kompensation der negativen Ladung der gitterhaften Alumokieselsäure 1 Na, bzw. auf 2 Al 1 Ca notwendig [!].

T. F. W. B.

Geological Explorations in Western Tibet. By E. NORIN. Octavo. Pp. 214; figs. 34; pls. 22; colored geological map in pocket.

There are few regions of the earth comparable to the vast, desolate, desert highland of the Tibetan Plateau, still geologically a terra incognita.

Professor Norin, together with Dr. N. Ambolt, started a geodetic and geological reconnaissance of western Tibet and the adjoining parts of K'un-lun in 1932 and was able to carry out an extensive scientific program in spite of great hardships and tribulations in this most inhospitable region, which is devoid of almost everything necessary for the maintenance of both man and beast and is fringed by hostile tribes. The members of the expedition were twice captured and kept imprisoned, once by Chinese soldiers, once by armed Tibetans.

The carefully prepared report is notable for its lucidity and its accurate and quantitative statements. Excellent descriptions of the geology of the areas are given; but the author goes far beyond a mere description; discussions, deductions, and comparisons with other areas are generously incorporated.

The last chapter is a treatise on the magmatism in Karakoram and western Chang-Thang, based on extensive laboratory studies of a great number of rock samples: ten new chemical rock analyses are given. Furthermore, geometrical analyses, optical data, and extended petrographical calculations based on new and old data are used for the construction of variation diagrams and other graphs illustrating petrogenetic relations.

Also included is a geological map in colors of western Chang-Thang and adjacent parts of K'un-lun and Karakoram, on a scale of 1:750,000, and covering the area between 34°–38° N. and 76°–82° E.

The book is written in English but has the following appendixes printed in German: (A) "Jungpaläozoische Korallen von Sven Hedins Zentralasien-Expedition 1932," von Franz Heritsch; (B) "Permische Bryozoen aus Nordtibet," von Dr. Karl Metz, Leoben; and (C) "Mesozoische Fossilien aus dem Tschiptschaktal und Lingschi-Thang," von Hans Frebold.

T. F. W. B.

Festschrift Paul Niggli, zu seinen 60. Geburtstag. (Herausgegeben von seinen Schülern, Mittarbeitern und Fachkollegen an den schweizersichen Hochschulen.) Zurich: Gebr. Leemann & Co., 1948. Pp. 568, with photo of Professor Niggli.

This volume contains forty-seven scientific papers (which form Heft 1, Band 28, of the "Schweizerische mineralogische und petrographische Mitteilungen") and a complete bibliography of the publications written by Professor Niggli (from 1917 through 1948), as well as the Ph.D. dissertations completed under his supervision.

It is impossible within the allotted space to review individually any of the forty-seven papers. Because all papers are written by Niggli's pupils and co-workers, the complete volume reflects the extraordinary breadth of knowledge spanned by the intellect of Professor Niggli: mathematical and physical crystallography, oil geology, geochemistry, petrography, petrogenesis, paleontology, ore geology, crystal chemistry, sedimentology, petrofabrics, soil mechanics, geophysics, and much more. Indeed, this volume is a shining manifestation of the high scientific standard, vigor, and dynamic activity of the Niggli school in Swiss geology.

T. F. W. B.

Technische Gesteinskunde ("Technologic Aspects of Rocks"). By A. VON MOOS and F. DE QUERVAIN. ("Lehrbücher und Monographien aus dem Gebiete der exakten Wissenschaften," No. 15, "Mineralogisch-geotechnische Reihe," Band 1.) Basel: Verlag Birkhäuser, 1948. Pp. vii+221; figs. 115. Fr. 29 (unbound); 33 (bound).

This book, written by two staff members of the Zurich Institute of Technology, is a useful outline for engineering geologists, especially those who deal with problems of tunnellng operations, foundations, highway construction, and testing laboratories.

Throughout, those properties of minerals and rocks are stressed that need special attention by engineers. Descriptions are brief but up to date and complete; many determinative and testing methods are illustrated; facts and figures are compiled in numerous tables; and references to technical papers are given in footnotes. Comprehensive works in the field of petrography, mineralogy, engineering geology, ceramics, and soil mechanics are cited in three pages at the end of the book and are followed by an index.

R. B.

The "Stratigraph." Los Angeles 27, Calif.: Mohave Sales, Inc., 1948. $7.50.

The "Stratigraph" is a sheet of flexible plastic, $10\times7\frac{3}{4}$ inches, that enables geologists to determine rapidly and accurately certain unknown structural data of inclined strata or tabular rock bodies, from those determined at the surface or through drill records.

It is a square, composed of 100×100 small squares, and each 10×10 square is marked by a heavier line. Circles concentric with the upper left corner connect the 100 intersection points on the upper margin with those on the left margin. Around the upper left corner a plastic strip with ruled black line rotates across the sheet, and tick marks on the right and lower border of the square indicate in degrees the angle between the strip and the upper margin of the square. Thus, for any position of the strip, two series of right triangles exist on the sheet, the length of whose sides can be read off.

On the back, diagrams explain the use of the grid for twelve selected problems, such as thickness of beds from dip and depth; depth from surface distance and thickness; dip of beds from depth and thickness. The accuracy of this instrument approaches closely that of trigonometric solutions and, like the latter, is more than adequate for most geological examples. For speedy answers this is a convenient tool. It is sold in a leather envelope.

R. B.

Heat Conduction, with Engineering and Geological Applications. By L. R. INGERSOLL, O. J. ZOBEL, and A. C. INGERSOLL. ("International Series in Pure and Applied Physics.") New York: McGraw-Hill Book Co., Inc., 1948. Pp. xii+278; figs. 42.

This volume is modeled after the well-known text, *Mathematical Theory of Heat Conduction* (Ginn & Co., 1913), by the senior authors, but the material has been thoroughly revised; much new material has been added, including many useful tables and an extensive bibliography; the nomenclature has been modernized; but the mathematical treatment is essentially the same as in the earlier work.

The development and explanation of heat-conduction equations of many types is, for the most part, detailed enough to be followed with a minimum of difficulty by readers moderately familiar with calculus; ellipsis is evident chiefly in the occasional differentiation of involved functions—the authors seem to make the erroneous but flattering assumption that, although transformations and developments of equations involving integrals may be somewhat difficult, all readers have the art of differentiation at their fingertips. However, the use of the important basic equations is so well illustrated by fully worked-out examples that a geologist or engineer should be able to use the equations whether the derivation is clear or obscure; but one would miss much of the fun of mastering a problem if unable to follow the explanation of the general case. The writers are to be congratulated for writing a book on a highly mathematical subject that is so generally readable and that should be understandable by anyone with the equivalent of a one-year course in calculus. This reader, however, found the frequent use of "while" (properly a time term) for an adversative conjunction, where "although" would better carry the meaning, to be singularly unfelicitous in a book primarily concerned with time relations.

Chapter i ("Introduction") has been much improved; the tables of nomenclature and conversion factors and the expanded discussion of the application of conduction-equation mathematics to fields other than heat conduction are helpful improvements; the use of dimensional formulas could well have been preceded by a brief explanation and followed by a reference to their use in modern theory.

Chapter ii ("The Fourier Conduction Equation") contains the same material found in chapter ii of the earlier book. It presents a clear but concise derivation of the Fourier equation, together with a brief section on the use of boundary conditions in determining particular solutions of the general equations.

Chapter iii ("Steady State, One Dimension") illustrates the plan used in most of the remaining chapters of the book. The problem to be studied is first explained, the general form of the Fourier differential equation that applies is presented, and the boundary conditions that must be satisfied are enumerated. This is followed by a brief, but adequately explained, derivation of an algebraic or integral equation that will allow arithmetical substitutions, satisfy the boundary conditions, and be a particular solution of the Fourier equation. After treating the general case of one-dimensional heat flow, a section is devoted to illustrative applications of the equations derived in the preceding section, and the chapter closes with a half-dozen problems, with the answers given, for the student to work out and check. The examples chosen in this chapter are of interest to engineers rather than to geologists and deal chiefly with problems of insulation and contact resistance.

Chapter iv ("Steady State, More Than One Dimension") is also concerned with insulation and cooling problems of engineering interest, but section 4.10, "Subterranean Temperature Sinks and Power Developments; Geysers," contains some material for the geologist.

Chapter v ("Periodic Flow of Heat in One Dimension") treats several geologic problems; heat waves—cold waves, periodic heat flow and climates, and thermal stresses all suggest possible applications to our field.

Chapter vi ("Fourier Series") is identical with chapter vi of the earlier volume; it is devoted to the step-by-step derivation of the Fourier sine and cosine series that must be used in many of the particular solutions of the Fourier heat-conduction equation, where unsteady-state heat-flow problems are to be solved. It is a pleasure to see such a generally readable explanation of a somewhat difficult subject, but it is regrettable that the occasional ellipsis present in the earlier work has not been eliminated.

Chapters vii and viii ("Linear Flow of Heat, I and II") take up many problems of heat flow in one dimension (taken as parallel to the x-direction). Subjects of interest to geologists will be found under several headings: the "concrete wall," mathematically identical with the problem of a cooling dike, also covered by problem 4 (p. 85) (intrusion of "a slab of molten lava"); the heating or cooling of a semi-infinite body whose surface is held at constant temperature; temperature of contact of thermally dissimilar bodies; the thawing of frozen soil; the cooling of a lava sheet; the cooling of the earth, assuming an initially hot homogeneous mass—both with and without radioactive heat sources; the temperature in rock decomposing exothermically; postglacial climate calculations from temperature gradient measurements; and the heating or cooling of slabs suddenly immersed in a medium of different temperature (theory of the fireproof wall).

The mathematical discussion of the cooling of the earth, on the assumption that a quarter of the heat loss is due to radioactivity, gives a figure of 45,000,000 years for its age, as compared with 22,000,000 years for a nonradioactive earth; the concluding paragraph of this section unfortunately states that these estimates are "of the same order of magnitude as those arrived at from geological considerations, such as stratigraphy, sodium denudation, etc., . . . although far short of the 2×10^9 years which represents the present trend of thought." As no explanation of the discrepancy is offered, some engineering readers will be misled, and geologists may feel baffled and disturbed. In the section on temperature in decomposing granite, references should have been cited for the statements (p. 115) that "areas of granite undergoing decomposition are several degrees warmer than surrounding rock" and that granite gives out heat during decomposition of the order of 100 cal/gm.

Chapter ix ("Flow of Heat in More Than One Dimension, Unsteady State") includes such geologic applications as the cooling of a laccolith, heat sources and heat sinks as related to geysers, the general cases for the cooling of a rectangular parallelopiped and of a finite cylinder in material of the same sort, and the use of conduction equations in such liquid-diffusion problems as the drying of porous solids. A helpful addition to this part of the book is the table on pages 184–185, which summarizes the heat-conduction equations for many differently shaped bodies under specified conditions.

Chapter x ("Formation of Ice") is essentially the last chapter of the earlier text and gives both Neumann's and Stefan's solution of the problem of ice formation for one-dimensional heat flow involving latent heat of fusion in a semi-infinite body with its surface at constant temperature.

Chapter xi ("Auxiliary Methods of Treating Heat Conduction Problems") gives several approximation methods of solving heat-conduction problems, a few of which may take less time than the analytical methods which they replace. Some, however, can be used to advantage where the problem is too complicated to be handled readily by classical mathematics. The explanation of the step method, as illustrated in the problems of ice formation about pipes, and the warming of soils, will be found helpful by geologists in a variety of problems. The brief section on electrical methods of solving heat problems is good, but it is to be regretted that the use of thermal models is entirely neglected.

Chapter xii ("Methods of Measuring Thermal Conductivity Constants") concludes the body of the text with a brief description of several methods of determining *k*. It is followed by thirteen appendixes: thermal constants, an extensive bibliography, and many mathematical tables especially useful in the solution of heat-conduction equations are given. A good index completes this excellent work.

The book is well illustrated with diagrams and graphs, the print is clear and easily read, the paper is good, and the text seems singularly free from typographical errors—only two minor ones were found by this reviewer in a cover-to-cover perusal of the volume. The wealth of material of diverse geologic interest and the lucid presentation of analytical methods should make it a most valuable and welcome addition to any geologist's library.

T. S. Lovering

COMMUNICATIONS AND ANNOUNCEMENTS

CAMBRIAN STRATA ON THE SOUTHERN SHORE OF DEVON ISLAND IN THE ARCTIC ARCHIPELAGO[1]

The southeastern part of Devon Island consists of pre-Cambrian rocks unconformably overlain by almost horizontal strata heretofore designated Silurian.

A geological reconnaissance of the Dundas Harbour area during the summer of 1948 by R. P. Nickelsen and D. B. Wales revealed that at least 700 feet of fossiliferous Cambrian limestones are present on the western side of Dundas Harbour.

The harbor extends inland 3.5 miles and is 1 mile wide at its entrance. A cliff approximately 2,400 feet high rises from sea level along the west side of the harbor. The lower 1,700 feet of the cliff consist of pre-Cambrian gneiss intruded by several diabase dikes. The upper 700 feet consist of dolomitic, arenaceous, and argillaceous limestones. The surfaces of the limestones are intensely fractured. Limestone blocks form talus in many places and cover the upper part of the pre-Cambrian rocks. Brownish dolomitic limestones, from within 20 feet of the unconformity, contain scattered brachiopod shells identified by Dr. W. C. Bell as *Micromitra*. The presence of *Micromitra* suggests strongly that the basal beds of the limestone series are Middle Cambrian. Approximately 400 feet above the unconformity, a thin-bedded limestone talus block yielded numerous trilobite glabella of the genus *Elrathia*. These also indicate a probably Middle Cambrian age.

The opportunity to study the rocks of the Dundas Harbour area was made possible by the United States Weather Bureau. Special acknowledgments are due to Professor R. L. Nichols, of Tufts College, and Mr. Charles Hubbard, chief of the Arctic Section, United States Weather Bureau, for aid in the field work, and to Mr. Vincent E. Kurtz, of Dartmouth College, who aided in the identification of the fossils.

Tentative plans for the summer of 1949 include a detailed study of the stratigraphy of the Dundas Harbour area to obtain information of the thickness and age relationships of the lower Paleozoic rocks of this interesting and practically unstudied area.

DONALD B. WALES

DARTMOUTH COLLEGE

REPRINTING OF KRYNINE ARTICLE

"The Megascopic Study and Field Classification of Sedimentary Rocks," by Paul D. Krynine (originally published in the *Journal of Geology*, vol. 56, March, 1948, pp. 130–165; 4 tables; 15 figures), the stock of which has been exhausted, has now been reprinted as Technical Paper 130, and is available at 25 cents per copy. The limit will be 50 copies per institution or organization.

Orders should be sent to:

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PENNSYLVANIA GEOLOGISTS' FIELD CONFERENCE

The Field Conference of Pennsylvania Geologists will hold its annual meeting May 27–29 at Lancaster, Pennsylvania. The department of geology of Franklin and Marshall College will be host. The major feature of this year's field excursion will be a trip through southern Lancaster County to study the Martic overthrust problems under the leadership of Professor Ernst Cloos, of the Johns Hopkins University. Other field excursions will include the old Wood chrome mine, Gap nickel mine, other old metal mines, and the flood plain of the Susquehanna River.

For further details, address Dr. Richard M. Foose, Department of Geology, Franklin and Marshall College, Lancaster, Pennsylvania.

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[1] Manuscript received February 7, 1949.

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STRUCTURAL PETROLOGY OF PLANES OF LIQUID INCLUSIONS[1]

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ABSTRACT

Planes of liquid inclusions are common in the quartz of igneous and metamorphic rocks. The planes have a high degree of preferred orientation, and it is believed that they represent fractures in the quartz grains that have been filled with intergranular liquid, which subsequently changed to individual inclusions by differential solution and deposition of silica. The planes of inclusions commonly extend through several quartz grains of different orientation with no deflection at the grain boundaries, indicating that there is little or no crystallographic control of their orientation. They have been found to have a remarkably uniform orientation over an area of 15 × 20 miles.

Detailed work on drag folds indicates that the planes of inclusions may develop in the shear directions of the deforming stress pattern and that one set tends to predominate. The poles to the planes tend to lie in a partial girdle about the fold axis. Some evidence was found indicating that the planes may develop in the tensional direction.

Quartz, then, may fail by shear or tensional rupture and in certain cases by deformation, which produces lamellae that may be a type of translation-gliding. Failure by rupture along crystallographic directions has been described by one investigator and has been produced experimentally in the laboratory. These different phenomena are believed to represent failure under varying environmental conditions, and the optimum development of any type of failure depends on such factors as confining pressure, temperature, composition, and the presence (and character) of solutions.

If several sets of planes of liquid inclusions develop at different times, the relative ages of the planes can be discerned from the character of the planes. Planes of inclusions that are very "young" appear as extremely thin sheets of liquid, whereas "old" planes are made up of inclusions that are nearly equant, with some of the inclusions having crystal planes as boundaries (negative crystals).

INTRODUCTION

GENERAL

Liquid inclusions have been recognized in the quartz grains of rocks since the advent of the microscope, and many observers have noted that they are commonly aligned in planes. The writer recently became interested in these structures when studying liquid inclusions in granitic rocks and associated quartz veins. Examination of oriented sections of a granitic rock from the Washington, D.C., area revealed a remarkable parallelism of the planes. Further observations on crystalline rocks in general showed that planes of liquid inclusions are very common; indeed, it is unusual to find quartz-bearing rocks that do not contain them. A statistical study of the orientation of the planes of liquid inclusions made by using the universal stage disclosed a remarkably high degree of orientation.

A cursory examination of the literature

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[1] This paper was presented in partial fulfilment of the requirements for the degree of Doctor of Philosophy at the Massachusetts Institute of Technology. Manuscript received June 29, 1948.

Linlithgow Library.
Imperial A[illegible]ltural R[illegible]rch Institute

left much to be desired in the way of explanation of the origin and significance of the planes of inclusions. Several conflicting theories had been advanced for their origin, and the high degree of orientation was not mentioned. It was at this stage that the writer decided to make an extensive study of the inclusions. The following report deals with those studies, which began in 1941 and continued until the summer of 1942, when full-time work was temporarily discontinued. However, laboratory work was continued intermittently, much of it at the Geophysical Laboratory, until 1946, when the present work was completed. Field studies were commenced and were continued, along with the laboratory studies, during parts of 1943, 1944, and 1945.

CHARACTER OF LIQUID INCLUSIONS

Liquid inclusions may be divided into two genetic groups: (1) primary and (2) secondary. Primary inclusions are emplaced during the growth of the host crystal, whereas secondary inclusions are introduced at some subsequent time. Only the latter will be considered here, as primary inclusions are usually randomly distributed throughout a crystal, or, if aligned in planes, the planes represent growth directions (zoning or lineage [Buerger, 1932]) in the crystal.

Liquid inclusions in quartz crystals vary in size from submicroscopic to cavities containing more than 50 cc. of solution. The large inclusions are, in all probability, primary inclusions emplaced during growth. The secondary inclusions which are considered in detail here rarely contain as much as 0.001 cc. of solution. They vary in size from submicroscopic up to a few hundredths of a millimeter in dimensions.

The shape of liquid inclusions is extremely variable. There are all gradations in shape from thin sheetlike inclusions with extremely irregular outlines to nearly equant inclusions bounded by crystal planes characteristic of the host crystal. Between these two extremes round, elliptical, cylindrical, cigar-shaped, pear-shaped, or irregular branching forms are found. Secondary inclusions usually have two dimensions considerably greater than the third. The inclusions usually lie in a plane with the planar element of the inclusions parallel to the plane.

Liquid inclusions are characterized by a vapor bubble which formed on cooling as a result of the much higher coefficient of expansion of the liquid than of the host crystal. In comparison with the liquid, the bubble is usually small, and it is not unusual to find it in constant motion (Brownian movement). This mobility of the bubble serves to distinguish liquid inclusions from glass inclusions found in some extrusive rocks.

If the fluid escapes from the inclusions (as always happens to those inclusions cut by the surfaces of the thin section), the liquid is replaced by air, which has a much lower index of refraction than has

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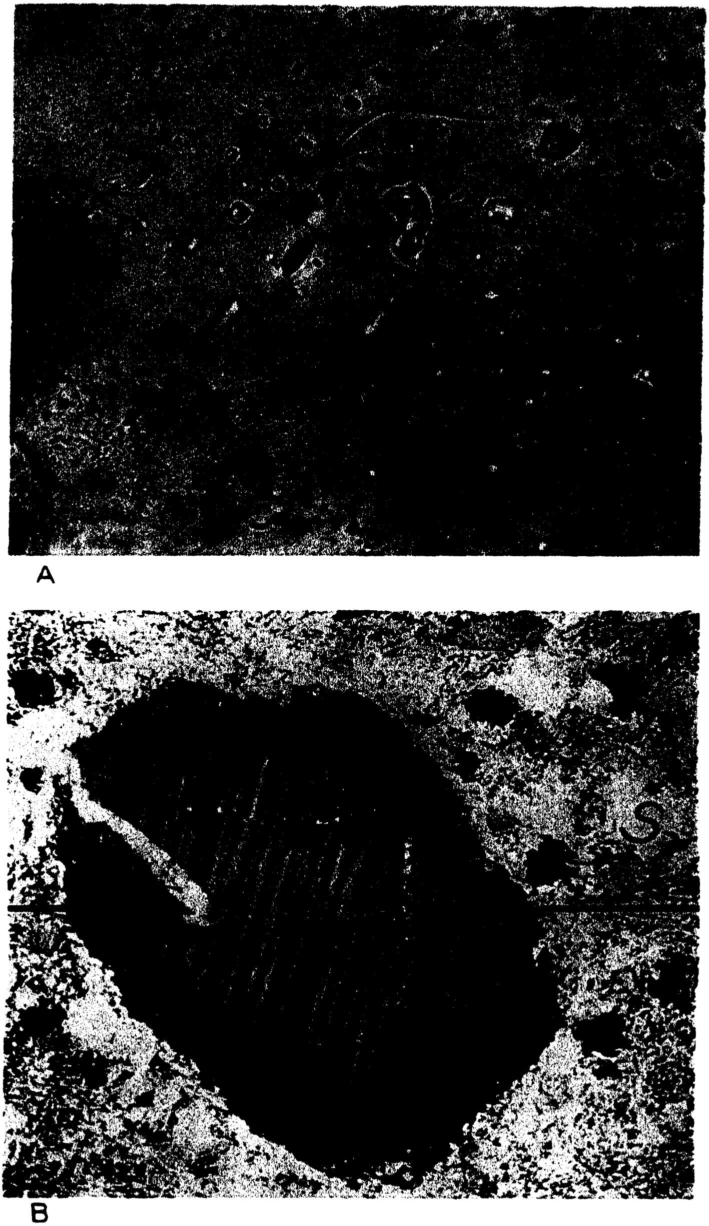
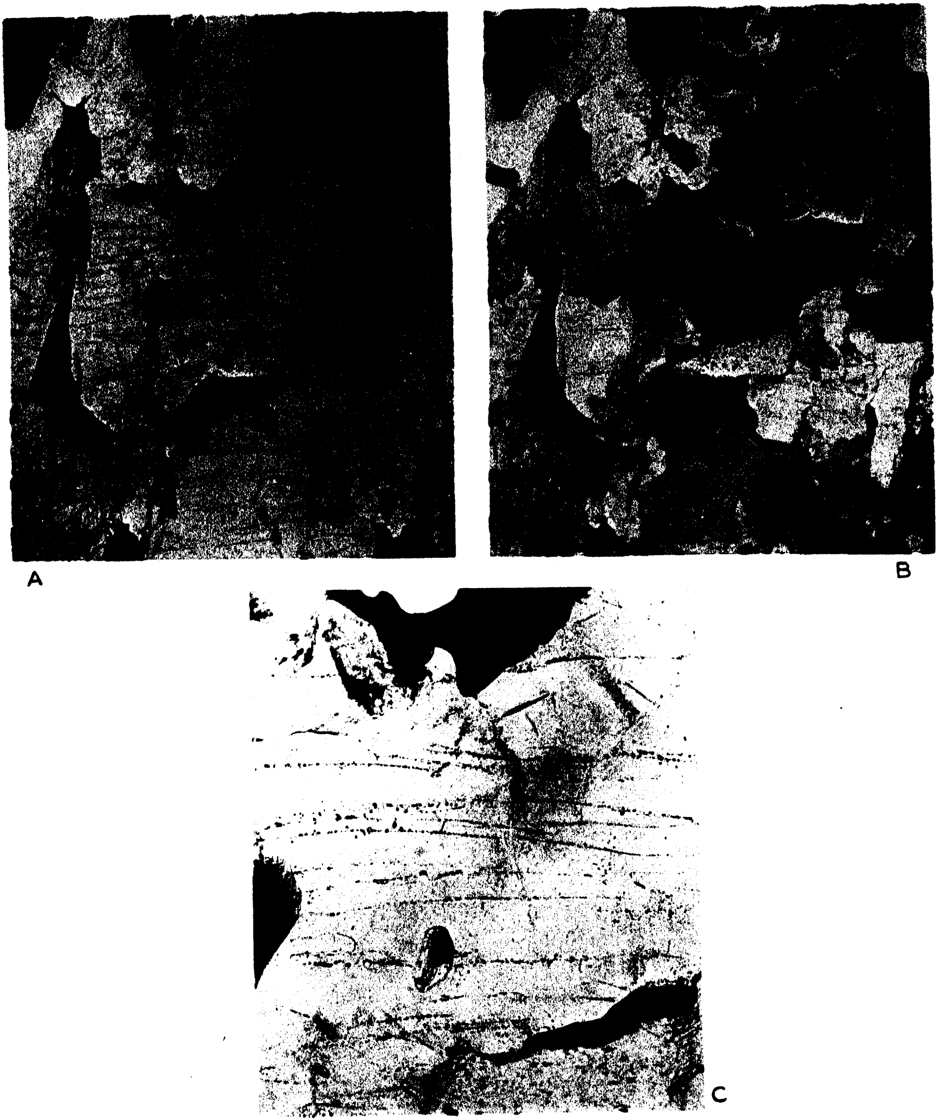
PLATE 1

A, Plane of liquid inclusions in quartz grain. The plane is nearly perpendicular to the microscope axis. The inclusions in the upper left-hand part of the photograph have boundaries that suggest imperfect negative crystal faces. ×500.

B, Quartz grain from Ajibik quartzite showing numerous lamellae (photomicrograph by Fairbairn, 1942).



Photomicrographs of quartz grains showing liquid inclusions (upper) and numerous lamellae (lower)



Photomicrographs of thin section of gneiss the host crystal. In many cases this index difference causes the cavities to appear opaque; in others it is obvious that the inclusion is filled with air. This apparently explains the common description in the literature of "rows of vapor, fluid, and opaque inclusions."

Liquid inclusions occasionally consist of two immiscible liquids and a vapor bubble. Gentle heating of this type of inclusion above 31° C. causes one liquid to disappear, indicating that it is carbon dioxide. Cubes of sodium chloride and potassium chloride are occasionally seen in liquid inclusions.

Plate 1, *A*, is a photomicrograph of a plane of secondary liquid inclusions. The plane is perpendicular to the microscope axis. The inclusions in the upper left-hand part of the photograph have boundaries that suggest imperfect negative crystal faces (see p. 334 for discussion of this feature).

ORIGIN OF SECONDARY PLANES OF LIQUID INCLUSIONS

Three different hypotheses have been proposed by various writers for the origin of planes of liquid inclusions. Two of the hypotheses require a secondary origin for the planes, whereas the third requires that the planes be primary. They may be termed (1) fracture, (2) solution, and (3) growth.

The fracture hypothesis was apparently first proposed by Hicks (1884). In describing planes of liquid inclusions in quartz grains of a conglomerate, he noted that they passed from one grain to another of different orientation without deflection. This led him to believe that they were secondary, that is, fractures which were "closed up by a secondary deposition of quartz."

The solution hypothesis is essentially chemical and was first proposed by Judd (1886). He believed that the planes of inclusions were formed by solution entering along "planes of easy solubility." These solution planes, writes Judd (p. 82), ". . . like the cleavage planes and glide planes, have definite relations with the symmetry of the particular system to which the crystal belongs." It follows, then, that planes formed in this manner should have definite crystallographic relations to the host crystal. This, of course, is not the case, as even Judd recognized, for later (p. 90) he writes: ". . . streams of bubbles are seen running through all the grains of quartz, quite irrespective of their orientation."

The third hypothesis for the origin of the planes of inclusions was first advanced by Dale (1923). The exact mechanism is not clear, but it is apparently a growth phenomenon, as he writes (p. 25): ". . . during the consolidation of granite, when the quartz . . . was crystallizing, the fluidal cavities were formed. . . ." Dale also recognized that some planes of inclusions were secondary, but he believed that those in granites were largely primary.

PLATE 2

A, Photomicrograph of thin-section of gneiss showing planes of liquid inclusions as lines or rows of dark specks in quartz grains. ×30.

B, Same as plate 2, *A*, with crossed nicols. The planes of liquid inclusions can be seen to pass from one grain to another of different orientation without deflection at the boundary. ×30.

C, Center portion of plate 2, *A*, at higher magnification. The planes of liquid inclusions are nearly vertical, hence they appear as lines. The undulatory nature of the planes can be seen here. Some individual inclusions can be seen along each plane at this magnification. ×100.

The investigation described in the following pages shows that planes of liquid inclusions in quartz grains of crystalline rocks have a high degree of orientation. The orientation in general is as pronounced as, say, the orientation of mica crystals in a well-foliated rock. This immediately suggests that any theory explaining their origin must include a process for aligning the planes into subparallel orientation over wide areas. The author's field work in the District of Columbia area shows an orientation persistent over an area of 200 square miles which therefore requires a regional deformation. It appears, then, that planes of liquid inclusions are a deformational structure of the same general type as cleavage, jointing, or faulting.

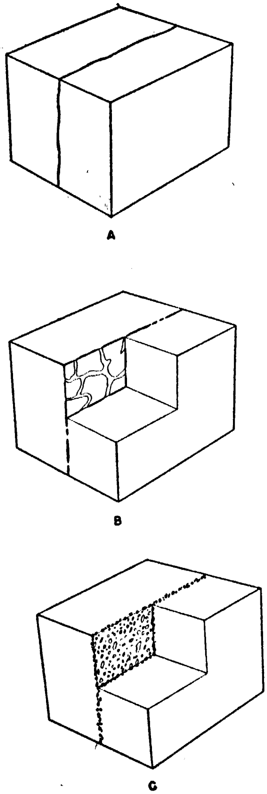


Fig. 1.—Schematic block diagrams illustrating the origin and development of a plane of liquid inclusions. *A*, the first stage in the development of a plane of liquid inclusions is believed to be fracturing of the quartz with subsequent filling of the fracture by intergranular or hydrothermal liquid. *B*, differential solution and deposition of silica have developed individual, thin, sheetlike inclusions. As the inclusions age, *C*, continued solution and deposition change the shape from thin sheetlike to more nearly equant. At this stage crystallographic planes appear as boundaries (see pl. 1, *A*) to the inclusions.

The author's theoretical mechanism for producing a single plane of liquid inclusions is illustrated in figure 1. The first step is fracturing of the quartz grains and filling of the fracture with liquid. Individual inclusions are then formed along the fracture by differential solution and deposition of silica.

The first stage in this final process of solution and deposition is the formation of large, thin, sheetlike inclusions having an extremely irregular outline. This type, when viewed in thin section on the universal stage, almost disappears when turned parallel to the microscope axis. Continued differential solution and deposition of silica gradually change the shape of the individual inclusions from thin sheets to more nearly equant bodies. The influence of crystallographic directions becomes apparent after the process has continued for some time, so that crystal planes appear as boundaries to the inclusions. At this stage the individual inclusions no longer lie in a sharp plane but have grown laterally to some extent, so that the plane has considerable thickness (*C*, fig. 1). Planes of liquid inclusions have been observed in all stages of development from a thin fracture filled with liquid to a broad band of nearly equant inclusions, many of which are bounded by crystal planes.

This suggests a method of estimating the relative ages of two or more sets of planes of liquid inclusions. The method can be illustrated in a mica schist in the

Washington, D.C., area, where an independent check is afforded by the relation of the two sets of liquid inclusions to small drag folds. The older set of planes is made up of nearly equant inclusions, whereas the younger set consists of thin, sheetlike inclusions. The older set of planes is not symmetrically related to the drag folds, and it has been established that these planes were rotated during the drag folding (fig. 77). On the other hand, the younger sheetlike planes are symmetrically related to the drag folds and are believed to be genetically related. The set of sheetlike planes would certainly have been considered younger without the independent evidence introduced by the drag folds.

The difference in age of the two sets of planes can be established by another character of the inclusions. This is illustrated by a thin section of a granite (locality unknown) containing two sets of planes, one of which has inclusions consisting of three phases: water, liquid carbon dioxide, and vapor bubble, whereas the other set consists of inclusions containing only water.

Solution and deposition along a fracture in a crystal can be impelled by mechanical or thermal energy. As differential pressures are certainly present, the so-called "Riecke principle" (1895) may be operative, and mechanical energy may motivate solution and deposition. This principle requires that, in the presence of saturated solution and differential pressure, solution occur at points on a crystal under greatest stress and that precipitation take place at points of least stress. Immediately after fracture, the two sides of the fracture would be in contact over a small proportion of the total area, and these portions of the crystal would be subjected to a greater stress than neighboring areas which are not in contact. Consequently, solution will take place in these highly stressed areas, and deposition will occur in the regions that are not in contact. The net effect of this process will be to increase the area under stress, thereby decreasing the stress per unit area. The area under stress would increase only by decreasing the area not stressed, and the end could give nearly equant liquid inclusions.

A second mechanism for differential solution and deposition would be important if the temperature of the crystal oscillated frequently. Increased temperature would cause silica to be taken into solution in the liquid of the inclusion (if the solubility increases with increasing temperature), and subsequent cooling would cause precipitation on the walls of the inclusion. It seems reasonable to assume that crystal faces would develop with continued solution and deposition, just as a growing crystal assumes euhedral faces.

Theoretically, it also appears reasonable that an irregular inclusion is not at its lowest possible energy level, and, given sufficient time—even without temperature oscillation—solution and deposition would take place until the inclusion has reached this lowest level. This would in all probability be expressed by a liquid inclusion bounded by crystal planes.

The Riecke principle would certainly be operative immediately after the fracturing of the quartz and would be responsible for initiating solution and deposition of silica. Later in the process the other two methods would in all probability be more important in continuing the solution and deposition.

METHOD OF STUDY

The orientation of the planes of liquid inclusions is measured in oriented thin sections in the conventional manner. The trace of each plane of inclusions to be measured is brought into coincidence with one of the cross-hairs by rotating the inner vertical axis of the universal stage. By rotation on the horizontal axis parallel to this trace, the position at which the plane of inclusions appears as a very fine line, and hence parallel to the microscope axis, can be determined with considerable accuracy.

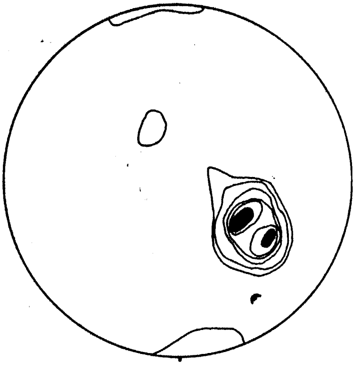


FIG. 2a.—Locality 19 (see fig. 25). 98 poles to planes of liquid inclusions measured by carefully aligning only those portions of each plane that are essentially planar. Contours 25–20–15–10–5–2–0 per cent.

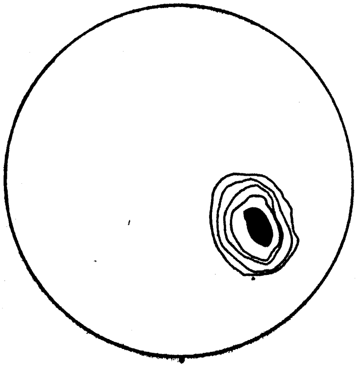


FIG. 2b.—Locality 19. 100 poles to planes of liquid inclusions from the same thin section as 2a. Measured by approximating the strike and dip of each plane. There is a slight loss in detail using this method which is similar to that used in measuring a rough joint or foliation plane in the field. Contours 25–20–15–10–5–0 per cent.

Identification and measurement of planes of liquid inclusions can be greatly expedited by preparing thin sections four or five times the normal thickness. Planes of large, equant liquid inclusions cannot be measured at all in a section of normal thickness. They appear as a row of inclusions which do not show a planar direction when rotated about a horizontal axis on the universal stage, thus accounting for the common description in the literature of "rows of liquid, gaseous, and opaque inclusions."

Planes of liquid inclusions are seldom uniform in the geometrical sense. The word "plane" is used here in the same sense as the geologist uses "fault plane" or "joint plane." The planes undulate considerably, and, when aligning the "strike" of a plane parallel to a cross-hair of the microscope, it is sometimes necessary merely to approximate the average "strike." When this question came up early in the investigation, several fabric diagrams of the planes of liquid inclusions were prepared from the same thin section, using slightly different techniques in aligning the planes of inclusions. First, a diagram was prepared by plotting only those portions of the planes that were actually planar. Another diagram from the same section, made by approximating the general strike of the planes in the microscope field, gave a similar diagram. Figure 2 illustrates the magnitude of the difference when the two methods are used. The differences illustrated indicate that some detail is lost by the second method of measurement; however, for the present reconnaissance work the second method is entirely adequate.

A more important source of error in this type of measurement is introduced through improper orientation of the thin sections with respect to the planes of inclusions. Figure 3 illustrates this error. In making the traverse *AA′*, nine planes of inclusions would be recorded, representing the vertical set of planes, with no recordings of the equally numerous horizontal planes. If the traverse were made along *BB′*, seven planes of each orientation would be recorded, this being the correct relation, provided that both sets of planes were perpendicular to the surface measured. If one set of the planes were nearly vertical while the other dipped steeply, neither traverse method would be correct (*CC′–DD′*).

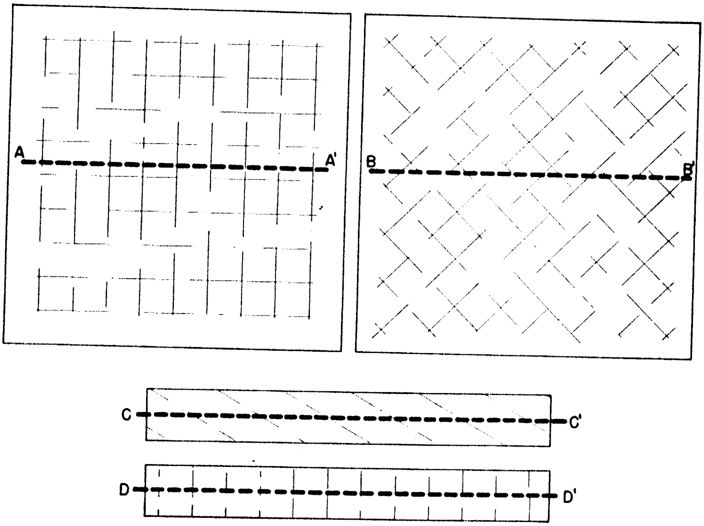


FIG. 3.—Schematic diagram illustrating some problems in traverse methods encountered when making statistical analysis of two or more sets of planes of liquid inclusions.

A mathematical expression can be derived for corrections where the thin section is not approximately perpendicular to the planes of inclusions. However, it was found to be much less time-consuming merely to cut sections perpendicular to each set of planes as determined in an initial section. Billings and Sharp (1937) have discussed similar problems encountered in making statistical studies of micas.

ORIENTATION IN INDIVIDUAL QUARTZ GRAINS

The most striking feature of these secondary planes of liquid inclusions in single quartz grains is the lack of crystallographic control of their orientation. This can readily be seen by crossing nicols and observing the planes extending across many quartz grains of different orientation with no deflection at the grain boundaries. A single plane within a quartz grain is usually nearly planar. However, the planes undulate considerably within a single grain, indicating, again, the lack of influence of crystallographic planes in the quartz.

Plate 2, *A* and *B*, illustrates the general character of the planes of liquid inclusions. The dark lines or rows of specks are planes of liquid inclusions. The planes are essentially vertical and appear to be continuous rather than made up of separate liquid inclusions. Plate 2, *C*, is a photomicrograph of the large grain in the center of plate 2, *A*. The undulating nature of the planes of liquid inclusions within a single grain can be seen at this magnification.

Figure 4 is a plot of the poles to planes of liquid inclusions in large single quartz grains. Where the planes were appreciably undulatory, several measurements were made at intervals along the plane, and the resulting plot of the poles connected by a line. These three diagrams were made from three different quartz grains in the same thin section. The lack of crystallographic control is well illustrated here, as the three *c*-axes of the grains have considerably different orientation, whereas the poles of the planes of liquid inclusions all have essentially the same orientation. This is further illustrated in figure 5, which was prepared by traversing the same thin section and plotting the *c*-axes of quartz grains together with the poles to planes of liquid inclusions. The high degree of preferred orientation of the planes of liquid inclusions with respect to the almost random orientation of the *c*-axes of the quartz grains further indicates the lack of control of the planes of liquid inclusions by crystallographic directions.

The planes of liquid inclusions in one quartz vein from the Washington, D.C., area are so unusual and informative that a detailed description and discussion is worth while. The usual shear planes of inclusions (see p. 347) locally deviate toward the direction of tensional rupture. A single plane of inclusions curves into short en echelon breaks, with the trend of the tensional breaks along the same direction as the original shear break (fig. 6). This resembles, on a small scale, shear faults (Newhouse, 1942) locally deviating toward the direction of tensional rupture. This has taken place *within* single quartz grains, indicating, perhaps, that the quartz behaved more like an isotropic material than like an anisotropic crystalline substance.

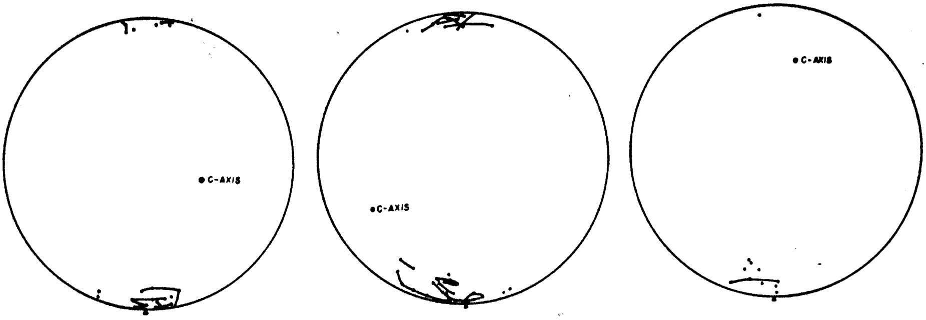


FIG. 4.—Orientation of planes of liquid inclusions in single quartz grains. Lines connecting poles illustrate the undulatory character of many of the planes. All three quartz grains are from the same thin section. The lack of crystallographic control of the orientation of the planes is apparent.

Planes of liquid inclusions vary greatly in their abundance in quartz grains. Thin sections of a specimen from locality 33 (see pl. 2, *A*, and fig. 27), for example, have planes spaced approximately at 0.06 mm., so that a traverse of 3 mm. perpendicular to the planes gave fifty measurements for a fabric diagram (fig. 59). This is one of the most abundant occurrences encountered. On the other hand, thin sections from locality 12 have only four or five planes that could be identified in completely traversing two large thin sections.

In general, planes of liquid inclusions appear to be most abundant in rocks containing a high percentage of quartz (i.e., quartzites). In some specimens showing a dearth of planes of liquid inclusions the foliation has nearly the same orientation as the planes elsewhere. It is suspected that movement which might have produced fractures that subsequently would become planes of inclusions was taken up by the mica, which was responsible for the foliation.

The average thin section from the Washington, D.C., area contains enough planes to prepare a representative fabric diagram on traversing the section four or five times.

In summary, the planes of liquid inclusions in individual quartz crystals show little or no relation to crystallographic directions in the quartz. In many respects the planes appear to have formed by rupture, with the directions of the rupture being controlled directly by the deforming stress uninfluenced by crystal structural planes within the quartz.

ORIENTATION IN SINGLE HAND SPECIMENS

One of the most striking features of fabric diagrams of planes of liquid inclusions is their high degree of orientation. Maxima up to 45 per cent in a 1 per cent area are not unusual, and 20 per cent maxima are very common. The diagrams look very much like mica diagrams from a well-foliated schist.

In the Washington, D.C., area, diagrams of planes of liquid inclusions commonly have only one strong maximum, as illustrated in figures 9 and 10. The planes of inclusions, then, usually occur in a single orientation, with only slight deviation from that orientation. Figures 11 and 12 are characterized by "split" maxima. This type has been found repeatedly in the Washington area (see p. 336). The genetic significance of these split maxima is not known. They may represent some slight crystallographic influence, or possibly they represent a slight change in the over-all stress relations at the time of their formation. If the latter, they should be characteristic of the area. It will be interesting to see what further work of this nature discloses elsewhere.

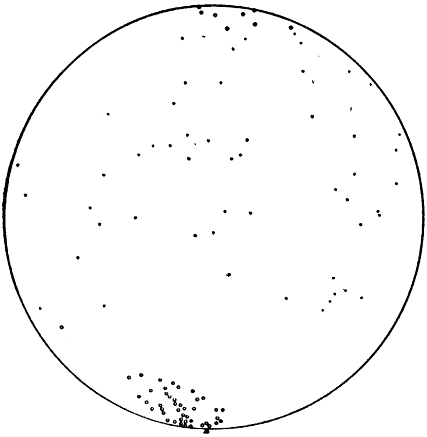

•-Poles to planes of liquid inclusions.
•-c-axes of quartz grains containing planes.

FIG. 5.—Diagram illustrating the relation between poles to planes of liquid inclusions and *c* axes of quartz. Prepared from the same thin section as fig. 4. The lack of crystal structure control of the orientation of the planes of liquid inclusions is illustrated here.

In contrast to the single strong maximum found in the foliated rocks of the area, two or more distinct maxima have been found in some quartz veins (figs. 13 and 14). The planes of inclusions responsible for the single maximum in the foliated rocks do not continue into the quartz veins. This might indicate that the quartz veins were introduced later than the fracturing that subsequently became planes of liquid inclusions (see p. 342).

Descriptions of planes of liquid inclusions from other areas frequently report "two sets of planes nearly at right angles." To check these observations, ten thin sections with a wide geographic distribution were selected and the orientation of planes of inclusions measured. The results are illustrated in figures 15–24, inclusive. As only one section was available in all cases, the diagrams are incomplete, in that other concentrations of poles to planes of inclusions may occur near the center of the diagrams (planar structures cannot be measured when nearly parallel to the section). Five of these ten sections show two sets of planes "nearly at right angles" (figs. 15–18, 22). Three of the diagrams (figs. 20, 21, 23) have maxima similar to those found in the Washington, D.C., area.

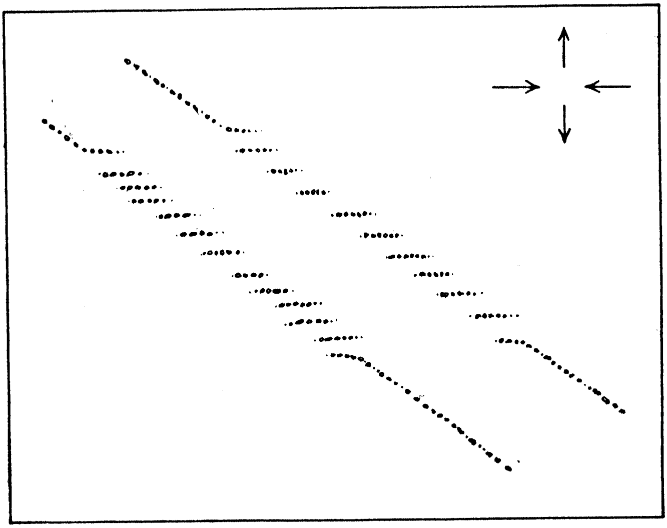
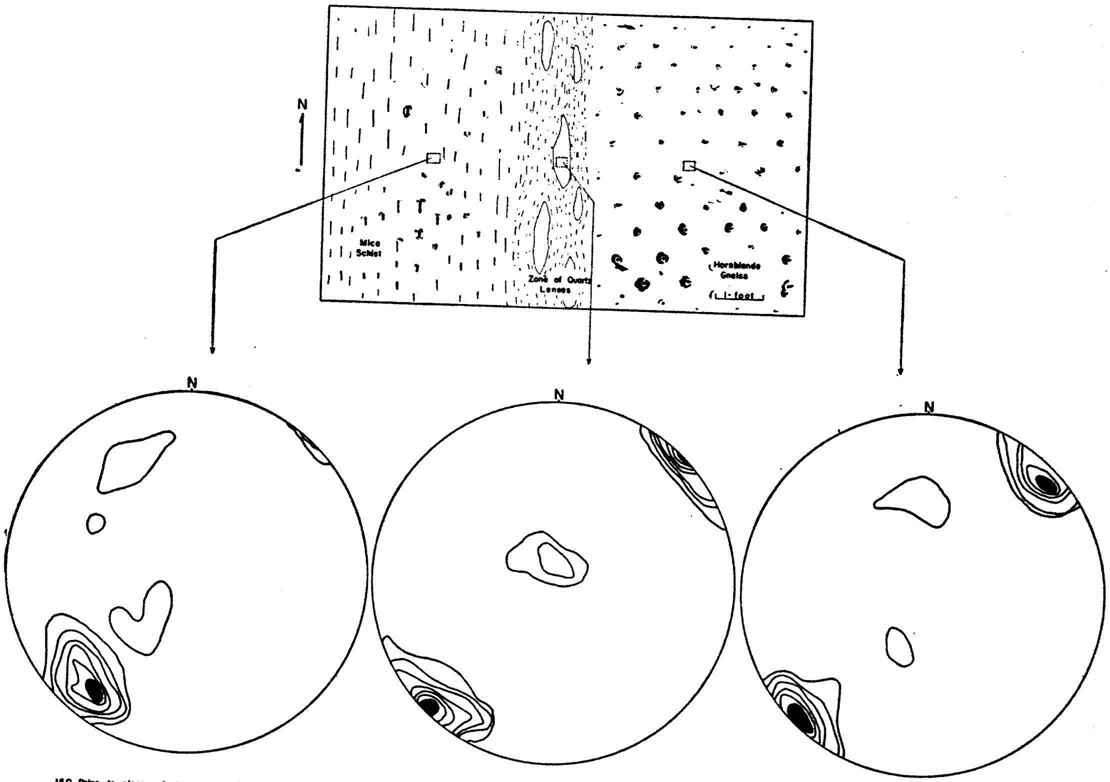


FIG. 6.—Sketch of the planes of liquid inclusions in a single quartz grain. The sketch is somewhat idealized. The planes of inclusions are perpendicular to the plane of the paper, and the rows of irregular circles and dots represent the trace of the planes on the surface sketched. The short en echelon planes are believed to be tensional breaks, as indicated by the arrows. The trend of the en echelon planes is believed to be the shear direction of the deforming stress pattern.

Many fabric diagrams of the poles to planes of liquid inclusions may have incomplete girdles. One specimen (fig. 24) has a complete girdle. It appears that the girdles occur about the fold axis, indicating that the planes are intersecting in the *b* fabric axis.

ORIENTATION IN SINGLE OUTCROPS

In a single outcrop, planes of liquid inclusions do not deviate to a measurable extent in hand specimens collected across the outcrop. Figure 7 illustrates this uniform orientation. (The slight difference in orientation of the maxima is within the experimental errors introduced during collecting, preparing the section, etc.) The planes cut across compositional boundaries with no deflection. For example, in figure 7 the planes have exactly the same orientation in the hornblende gneiss, the quartz lenses, and the



150 Poles to planes of liquid inclusions.
Contours- 25,20,15,10,5,2.

68 Poles to planes of liquid inclusions
Contours- (35-30),25,20,15,10,5,2.

98 Poles to planes of liquid inclusions
Contours- 25,20,15,10,5,2.

FIG. 7.—Sketch of an outcrop showing the orientation of planes of liquid inclusions in the different rock types

biotite schist. However, the number of planes per unit volume may vary considerably (see p. 339). A 1-cm. traverse across the planes in the quartz lenses may cross fifty planes, whereas only ten or fifteen may be encountered in a traverse of the same length in the biotite schist or the hornblende gneiss. This uniform orientation within the single outcrop has been found at numerous localities in the area studied.

In contrast with this uniform orientation in various foliated rocks, the quartz veins of the area present an entirely different problem. Figure 8 illustrates the orientation of the planes in two quartz veins and the enclosing mica schist. The planes of liquid inclusions in these veins show no obvious symmetrical relation to the planes in the schist. This suggests that the planes in the quartz veins are late and postdate the planes of inclusions in the surrounding rocks.

ORIENTATION IN WASHINGTON, D.C., AREA

AREAL GEOLOGY

The rocks of the area investigated are largely metamorphic and igneous, the only exception being a thin veneer of flat-lying, coastal-plain, sedimentary rocks in the southeastern portion. The metamorphic rocks are highly foliated with well-developed lineation. Lithologically the rocks are extremely diverse. This diversity is so pronounced that it seems certain that, if several different geological parties mapped the area, the resulting maps would show little, if any, uniformity of areal distribution of formational units.

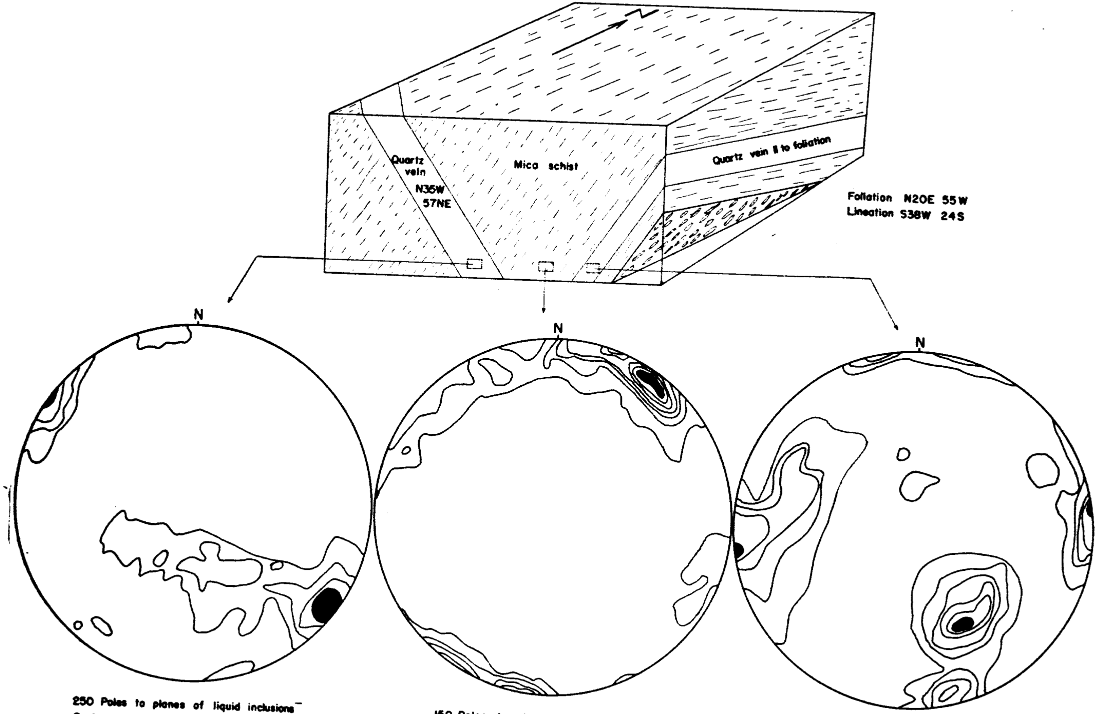
Most of the southern half of the metamorphic belt was mapped by Williams and Keith in 1893–99. The rocks are classified as Archean and the following types described: granite, granite gneiss, diorite, diorite gneiss, gabbro, metagabbro, metadiorite, serpentine, and soapstone.

Over 90 per cent of the area studied (fig. 27) is composed of rock types that were called "granite gneiss" or "diorite gneiss" by Williams and Keith (1901). These rocks are divided into two areas by a biotite granite which occurs in a narrow belt about 1 mile wide, running north-south through the entire region studied.

The granite and diorite gneisses are characterized by the almost universal presence of rock inclusions. The inclusions are usually lenticular, with their planar direction parallel to the foliation. They range in size from a few millimeters to a few meters in largest dimension. The smaller inclusions are usually quartz, although micaceous, chloritic, and pyritiferous inclusions are seen in places. The inclusions commonly have a bladelike habit, with the long axis parallel to the *b* lineation of the enclosing gneiss. The large inclusions commonly show structures that appear to be remnants of bedding or schistosity, which is usually parallel to the foliation of the enclosing gneiss. However, in some cases divergence from parallelism is pronounced.

These large inclusions would doubtless be called xenoliths by one disposed to consider this type of rock as intrusive igneous, whereas an advocate of granitization would probably interpret them as remnants of the original rock that had escaped the granitization.

The biotite granite is characterized by the absence of inclusions and by its uniformly light color. The biotite occurs in large wavy flakes that give the rock a pronounced speckled appearance. Foliation is not so pronounced as in the sur-



250 Poles to planes of liquid inclusions
Contours 9,7,5,3,1.

150 Poles to planes of liquid inclusions
Contours (16-14),12,10,8,6,4,2.

221 Poles to planes of liquid inclusions.
Contours 6,5,4,3,2,1.

FIG. 8.—Sketch of outcrop relations of mica schist and quartz veins, with fabric diagrams showing the orientation of planes of liquid inclusions. The quartz vein parallel to the foliation is approximately 1 foot thick where exposed, and the cross-cutting vein is approximately 2 feet in thickness.

rounding gneisses, except near the contacts and in the northern portion of the area. Lineation is poorly developed except in the northern part of the area.

STRUCTURAL FEATURES

Perhaps the most prominent structural feature of the region is the foliation (fig. 28). With the exception of a few minor areas, all the rock types encountered are well foliated. Nonfoliated granite was found in only two small areas. The foliation is so pronounced that one is inclined to classify all the rocks in the area as metamorphic rather than intrusive igneous. This tendency is enhanced by the scarcity of cross-cutting contacts. Although some contacts are discordant, in general they are parallel to the foliation, schistosity, or banding. The contacts are also parallel to the planar direction of the ubiquitous inclusions in the granite and diorite gneisses. Some of the more schistose horizons in the granite gneiss show small, close, contorted folds. The axial line of these folds is parallel to the regional lineation, and, in general, the axial planes appear parallel to the regional foliation. The "bedding" on the limbs of these small folds is also parallel to the regional foliation.

Locally the contact between, say, the granite gneiss and a schistose horizon in the granite gneiss may have discordant relations measured in feet. Embayments occur in the schist, and in some places blocks of it are apparently surrounded by the granite gneiss. Petrographic examination of the relations suggests a replacement origin for this type of contact, as fragments of the schist can be seen in thin section in all stages of digestion, with the smallest fragments showing unmistakable parallelism with the surrounding schist. If this were an intrusive contact, one would expect to find some fragments of the schist rotated somewhat by the "intruding" granite gneiss.

Foliation.—The most common type of foliation is a result of the alignment of mica flakes into subparallel orientation. The foliation is locally intensified by segregation of mica into layers alternating with quartzose layers. In some places banding due to alternating layers of amphibole and quartz-feldspar gives a pronounced foliation. In others the foliation is a result of alternate layering of quartz-feldspar and chlorite or of quartz and epidote. Foliation in the biotite granite is made conspicuous by the large, dark-colored mica flakes, but on close examination it is seen that it is in part due to segregation into bands of the quartz and feldspar. Thus, although the rock has the mineral composition of a granite, texturally it is a gneiss.

Lineation.—Five types of lineation have been observed in the Washington, D.C., area (fig. 29). Four of these are believed to be parallel to *b* (fold axis) and one perpendicular to *b* in the plane of foliation. The evidence that the four types are parallel to *b* is given by their parallelism with the axial lines of all the minor folds in the area (see above). The relation of these small folds to any major folds that may exist is not known. It is possible that the fold axes of the small folds may be either parallel or perpendicular to the regional fold axes. If they are parallel, the above-mentioned four types of lineation would be parallel to *b* of both the small folds and the regional folds. However, if the fold axes of the small folds are perpendicular to large-scale regional folds, then the four types of lineation would be in *b* of the small folds and *a* of the regional folds.

The four fold-axis types are: (1) Mica flakes in many of the rocks are so oriented that the poles to the cleavage have a

maximum at *c* and a girdle about the *b* fabric axis. This gives a pronounced linear structure on the foliation (*s*) plane parallel to *b* because the traces of mica plates on a surface parallel to *s* are statistically parallel to *b*. (2) This type is the result of orientation of elongated amphibole or tourmaline crystals and has been observed to be parallel to the fold axes of small folds throughout the area. (3) The third type is caused by pinching and swelling of alternate layers of different mineral composition (see p. 344). This is a gross lineation best seen at a considerable distance from the outcrop and is parallel to the fold axis of the small folds. (4) The fourth type is the result of parallel orientation of the long axes of elongate inclusions. These are also parallel to the fold axes of small folds.

A fifth type is seen in hand specimens as striae on cleavage surfaces. The striae are perpendicular to the four fold-axis types of lineation and hence are believed to lie in *a* of the small folds.

In addition to the above, several younger linear structures occur, with variable relations with the fold-axis types. For example, one schistose horizon in the granite gneiss commonly has numerous small drag folds (see p. 347) that are obviously later than the regional foliation. The fold axes of these small crenulations can be easily measured as a linear structure. Also, small but easily visible striae, which are perpendicular to the fold axes of these drag folds, can be seen at numerous localities.

RELATION OF PLANES OF LIQUID INCLUSIONS TO MAJOR STRUCTURES

The major structures as observed in the field are illustrated in figures 27–29. The foliation strikes predominantly north-south and dips steeply to the west. This is, perhaps, best illustrated in the compilation of all measured values shown in figure 25 where lineation measurements are shown by dashed lines and poles to foliation by solid lines. The linear structures shown in figure 25 all lie in the plane of the foliation, and all those values illustrated are believed to be parallel to the fold axes of minor folds seen in several localities in the area. Therefore, they are believed to represent lineation parallel to *b* of the small folds. Figure 25 is a compilation of one hundred and fifty-one measurements of strike and pitch of linear structures.

The linear structures generally pitch south to southwest in the eastern part of the area, whereas they pitch northwest in the western part of the area. The change from southwest to northwest pitch is gradual in the northern portion and surprisingly abrupt in the southern part. An explanation of the orientation of the linear structures will doubtless be obvious when the relations to similar structures over a much larger area are known.

The orientation of the planes of inclusions is illustrated in figure 30. Each orientation shown is the orientation of the maxima from a contoured fabric diagram. The planes of inclusions strike generally northwest and dip to the northeast. Figure 26 illustrates a compilation of all maxima, and it can be seen that they are highly oriented (maxima up to 23 per cent in a 1 per cent area) throughout the area. In fact, they are more uniform in orientation than either the foliation (16 per cent maxima) or the lineation (8 per cent maxima). The planes of liquid inclusions cut across the primary structures (foliation and lineation) and bear no symmetrical relation to them. This in itself would indicate that the planes of inclusions are later than the foliation and lineation and therefore are not genetically related.

It must be emphasized that any one of the maxima shown in figure 30 may be in error by as much as 20° because of the many manipulations that must be made in determining the orientation (i.e., orientation of hand specimen, orientation of thin sections, rotation of diagrams, etc.); however, in most cases it is believed that the errors are not more than ± 5°. It is believed that the orientation of the planes may be considerably more uniform than is illustrated because, when special precautions were taken to eliminate errors, no differences in orientation could be recognized between hand specimens from the same outcrop. This is illustrated in figure 7.

In considering the distribution of the planes of inclusions, it is remarkable that no macroscopic structures could be found that were related to the planes. One would think that an orogenic stress that could produce such a structure throughout an area of 200 square miles would manifest itself in some other way. After the orientation of the planes throughout the area was known, considerable time was spent in the field revisiting many outcrops in an effort to find some macroscopic structure that could be related to the planes. None was discovered. It is hoped that workers in other areas, where auxiliary structures may be developed, will be able to make some correlation.

The relation of the quartz orientation to the planes of inclusions is illustrated in figures 65–76. As might be expected, there is no consistent relation between the two, since the planes of inclusions merely indicate fracturing. Such an environment could hardly be expected to reorient the quartz. The quartz orientation is in most cases related to the primary foliation (*s*) and lineation (*b*). This again indicates that the planes of inclusions are later than the primary regional structures.

LATE STRUCTURAL HISTORY OF THE WASHINGTON, D.C., AREA

After the metamorphism that produced the foliation and quartz orientation, the entire region was subjected to orogenic movements that induced fracturing in the quartz grains of all the rock types. These fractures were strongly developed in a northwest-southeast direction, with a steep eastward dip. The fractures were filled with solution, either interstitial or hydrothermal, which subsequently became trapped as minute liquid inclusions in the quartz. The initial thin layer of liquid along each fracture became disseminated into numerous isolated inclusions by a process of solution and deposition of silica. This process of solution and deposition apparently continued for a relatively long period, since the liquid inclusions had begun to develop crystallographic planes as boundaries before the next orogenic disturbance. The next orogenic disturbance manifested itself as numerous small drag folds in the less competent horizons of the area. During the drag folding the quartz was locally reoriented in part and deformed subparallel to the shear directions of the deforming stress as deformation lamellae (see p. 348). The lamellae are believed to have formed while the folding was in process, as they occur as a partial girdle about the fold axis. After this period of deformation the quartz was again fractured locally, and again liquid filled the fractures. These latest fractures still remain as thin sheetlike liquid inclusions, indicating that they are relatively young compared to the other structural features described.

Sometime during the above-described

events quartz veins and pegmatites were introduced. They are, in part at least, later than the foliation; but otherwise no evidence has been discovered to fit them more closely into the sequence of structural events.

GEOLOGICAL SIGNIFICANCE OF PLANES OF LIQUID INCLUSIONS

GENERAL

Planes of liquid inclusions have not been identified megascopically, nor have they been related to any megascopic structures, such as joints and fractures. The commonly described *ac* and *bc* fractures, both of which can usually, if not always, be recognized in the field, have not been recognized in the Washington, D.C., area; therefore, it is not possible to compare them directly with the planes of liquid inclusions. The most important distinction between the fractures and planes of liquid inclusions is that the latter are confined to the quartz grains of the rock, whereas joints and fractures usually represent failure of all the minerals in the rock. It is possible that megascopic jointing and fracturing are a relatively shallow-zone manifestation of the same stress environment that produces planes of liquid inclusions at greater depths.

Heretofore the only suggestion of the orientation of the planes of liquid inclusions with respect to a direction in the deforming stress pattern has been the en echelon tension-like planes described on page 338 (see also fig. 6). These planes would seem to indicate that either the shear or the tensional direction in the deforming stress pattern may be favorable for the development of fractures that subsequently become planes of liquid inclusions. It has been mentioned (p. 340) that the poles to planes of liquid inclusions commonly lie in incomplete girdles. These may be perpendicular to the fold axes. The evidence for this belief, presented in the next section, is the relation of a set of extremely thin sheetlike planes of inclusions to small drag folds found in a schistose horizon in the granite gneiss.

PLANES OF LIQUID INCLUSIONS IN SMALL DRAG FOLDS

The drag folds are younger than the foliation, since the mica flakes are bent around the folds with no evidence of recrystallization. The planes of liquid inclusions previously described are rotated by this folding and therefore antedate it. The maximum in diagram *A* of figure 77 shows the orientation of these planes in the portion of the schist that has not been rotated. Only a few of these planes could be found in the rotated portion of the fold, but all were obviously rotated during the folding. These planes of inclusions are therefore older than the drag folds and are not genetically related. In addition to these planes, characterized by a broad plane of nearly equidimensional inclusions, another type is present in sections of the drag folds. This type is characterized by extremely thin planes of sheetlike inclusions. Diagrams *B* and *C* of figure 77 illustrate their orientation with respect to the drag folds. They are symmetrically related to the drag folds and are believed to be also genetically related. They are essentially in the shear direction of the conventional stress pattern. In diagram *B* of figure 77 most of the planes are perpendicular to the foliation, which is one of the shear directions of the drag fold. In diagram *C* of figure 77 most of the planes are again nearly perpendicular to the rotated foliation, but this is also a shear direction of the drag fold. In both *B* and *C* there are some planes of inclusions in both shear directions,.but one of the directions predominates in each case, hence providing a possible explanation of why only single-maximum diagrams are found in most specimens in the area. The drag fold is composed of alternating layers of mica and quartz, and in all probability the movement which theoretically should have produced two equally dense maxima in *B* of figure 77 was partially taken up by the less competent mica layers when the shear was along the plane of the mica layers. It should be noted also that a few planes were found that were in the tensional direction in the drag folds.

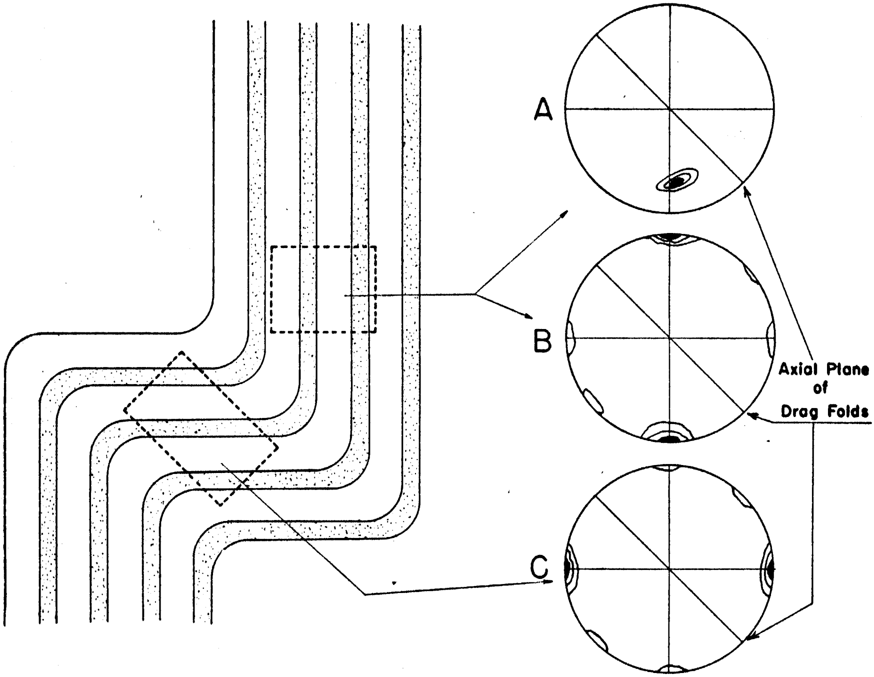


FIG. 77.—Diagrammatic cross section of drag fold, with schematic fabric diagrams illustrating the orientation of planes of liquid inclusions in the fold. *A*, poles to planes of liquid inclusions characterized by a broad plane of nearly equidimensional inclusions. *B–C*, poles to planes of liquid inclusions characterized by extremely thin planes of sheetlike inclusions.

Diagrams *B* and *C* are sketches of what is believed to be the correct interpretation of the orientation of the thin sheetlike planes. Conventional contoured diagrams were not prepared because only a few planes were found in the thin sections available. Measurement of these planes is extremely tedious because they can be seen only when nearly parallel to the microscope axis and only a few were found in carefully traversing several large thin sections.

RELATION BETWEEN DEFORMATION LAMELLAE AND PLANES OF LIQUID INCLUSIONS IN QUARTZ[2]

Deformation lamellae in the quartz grains of these drag folds have been described previously (Ingerson and Tuttle,

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[2] Since the manuscript for this paper was prepared, an excellent paper has been published in this *Journal* (November, 1947) by N. Allen Riley, describing deformation lamellae and planes of inclusions from the Baraboo quartzite. His results

1948). The lamellae were found to occur in the shear directions of the deforming stress pattern, and it was shown that the orientation of the lamellae is unrelated to definite crystallographic directions.

Lamellae in quartz were first described in detail by Böhm (1883), and it is interesting to note that Böhm believed the lamellae to be planes of inclusions. Where the inclusions could not be seen, Böhm suggested that they were merely ultramicroscopic in size. A somewhat different explanation has been offered by other observers (Kalkowsky, 1878; Ingerson and Tuttle, 1948; and Fairbairn, 1941). Lamellae are believed to be an entirely different type of structure from planes of inclusions, and the confusion has been brought about by the fact that planes of inclusions commonly occur with the lamellae. Lamellae (pl. 1, *B*) are planar structures manifested by a slight difference in refractive index between the lamella and the surrounding area. In some cases the birefringence serves to distinguish their presence. They can be seen in thin section only when they are nearly parallel to the axis of the microscope. In many cases it is necessary to have sections several times the normal thickness to recognize them. They are not definite breaks in the quartz grains and do not pass uninterrupted from one grain to another of different orientation as do planes of liquid inclusions. In general, lamellae are more common within a zone from 10° to 30° of the base, although they have been observed in all positions from 0° to 90° of the base.

Lamellae have been interpreted as translation-gliding planes by some observers (Mugge, 1896; Sander, 1930). Fairbairn (1941) suggested that a glide line (*m*:*r*) was operative rather than one or more glide planes. Ingerson and Tuttle (1948) concluded that the lamellae are not controlled by definite crystallographic planes or zones. They suggested that the quartz had failed more nearly like an isotropic material than like a crystalline substance.

The orientation of the lamellae in the quartz of the drag folds previously described is illustrated in figure 78. The relation between the orientation of these lamellae and the planes of liquid inclusions can be seen by comparing with figure 77. Both structures are related to the shear direction of the deforming stress. This suggests that the two structures are produced in similar stress environments. The deformation lamellae tend to be less highly oriented than the planes of inclusions, indicating, perhaps, that they were formed at an earlier stage while the quartz grains were still being rotated by the deforming stress. Rotation of individual quartz grains did not occur during or after the fracturing that produced the planes of inclusions because a single plane can be seen crossing many quartz grains of different orientation without deflection at the boundaries.

RELATIONS BETWEEN PLANES OF LIQUID INCLUSIONS AND THE RIFT, GRAIN, AND HARDWAY OF GRANITES

Planes of liquid inclusions have been reported as being parallel to the rift, grain, or hardway of granites. Some investigators have reported parallelism between planes of inclusions and one of these directions, and others between planes of inclusions and two of these directions. In all examples the relation between the two structures was apparently

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are essentially in agreement with this work; i.e., deformation lamellae and planes of liquid inclusions are not related to macroscopic structures but are related to one another. Planes of liquid inclusions were found to have single strong maxima, two maxima, and partial girdles.

investigated by inspection of thin sections without the aid of the universal stage, so that orientation of the planes could not be accurately determined. For example, figure 79 (Dale, 1923, p. 18) is a camera lucida drawing of a section parallel to the hardway of a granite and is intended to illustrate that the planes of inclusions are parallel to the rift and grain. It can be stated with certainty that neither set of planes of inclusions illustrated is parallel to the rift or grain of the granite. The "strike" of the two sets of planes with respect to the thin-section surface is nearly parallel to the rift and grain directions, but the "dip" in both sets obviously does not coincide with the rift or grain in either case.

The interpretation placed on this example throws doubt on other observations made without the universal stage and emphasizes the need for careful statistical study of such structures. For example, planes of liquid inclusions are reported to be parallel to the rift of some pre-Cambrian granites of Quebec (Osborn, 1935). However, the method of determining the orientation is not given, and the observation may be subject to the same misinterpretation as that mentioned above.

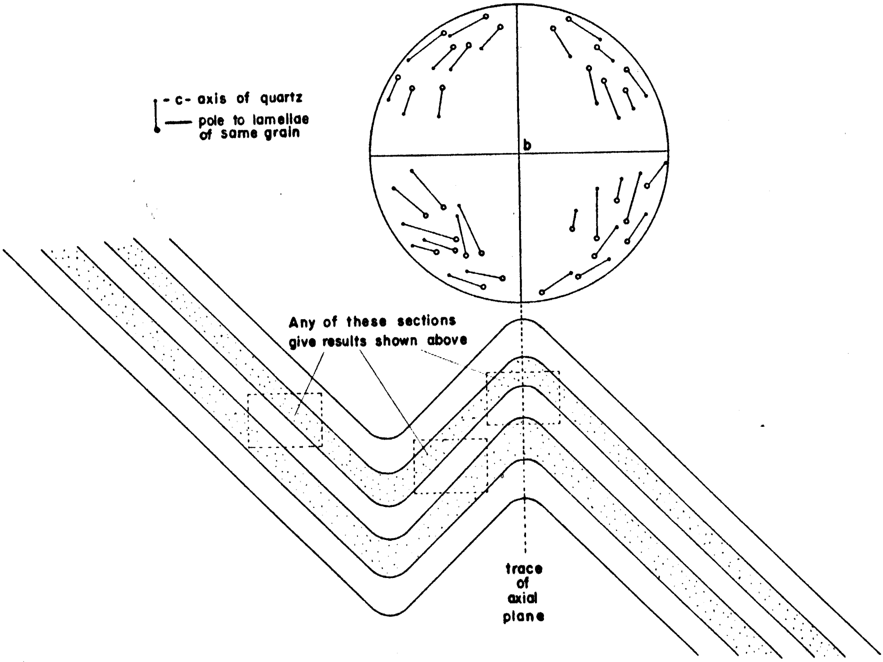
c- axis of quartz
pole to lamellae of same grain
b
Any of these sections give results shown above
trace of axial plane

FIG. 78.—Diagram illustrating the relation between *c* axes, poles to lamellae, and fold axis of a small drag fold in mica schist. Note that the lamellae have the same orientation as the planes of liquid inclusions in fig. 77.

RELATION BETWEEN SHEAR AND TENSION FRACTURES AND PLANES OF LIQUID INCLUSIONS

Sander (1930) has studied and described the orientation of four types of fractures in quartz. Two sets (*ac* and *bc*) are tension, and two sets (intersecting in *a* and *b*) are shear fractures. Unfortunately, detailed description of the fractures is unavailable to correlate them directly with planes of liquid inclusions. All four types are symmetrically related to the primary structures of the rocks and to the quartz orientation, whereas the planes of liquid inclusions considered here are not so related. At least two of the fracture types (*bc* and *b*) are megascopic structures, hence differing from the planes of inclusions in this respect. The four types of fractures have one important feature in common with planes of liquid inclusions, namely, the lack of control of their orientation by crystallographic directions in the quartz. This further emphasizes the isotropic nature of observable quartz failure in deformed rocks.

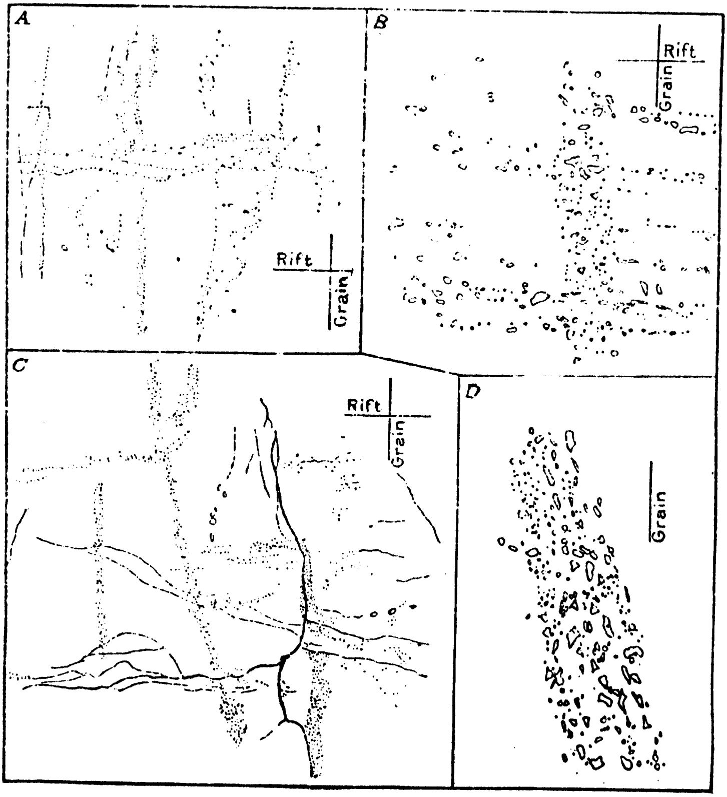

FIG. 79.—From Dale (1923): "Camera lucida drawings of quartz areas in an enlarged thin section of biotite granite from the Redstone quarry in Conway, N.H., cut parallel to the 'hard way.' *A*, enlarged 62½ diameters, shows sheets of cavities and incipient cracks in both rift and grain directions. *B*, from a different quartz area, enlarged 175 diameters, shows the variation in the size of the cavities and their arrangements in both rift and grain directions. *C*, from still another quartz area, enlarged 23½ diameters, shows conspicuous rift and grain cracks, filled with fibrous white mica, coinciding here and there with the sheets of cavities, also two parallel exceptional fractures crossing the rift and grain directions diagonally. *D* shows part of one of the grain sheets of cavities of *C*, enlarged 175 diameters in order to show the shapes of the cavities. As most of these sheets of cavities in the section appear to undulate in the direction of the line of vision when placed under the microscope the outlines of all the cavities become visible only by altering the focus."

ENVIRONMENT FOR PRODUCTION OF PLANES OF LIQUID INCLUSIONS

Planes of liquid inclusions have been described by Anderson (1945) as being related to crystallographic directions in quartz. The quartz studied by Anderson occurs in small veinlets in a quartzite and is believed to represent low-grade metamorphism, as it lacks secondary silica and contains much chlorite. The significance of this work is open to some question because only eight quartz grains were studied and statistical methods were not used. However, if it is assumed that the planes of liquid inclusions are related to crystallographic directions in some cases, then it seems reasonable to expect that a different stress environment is responsible for the control of fracturing by crystallographic planes in one case and the lack of control in another. The rocks in the Washington, D.C., area have been subjected to a very high-grade metamorphism and probably were buried to a greater depth than the low-rank chloritic rocks in the area investigated by Anderson. The speculation is therefore warranted that quartz may behave like an anisotropic material when deformed at relatively shallow depths, but at higher pressures (and temperatures?) and greater depths the crystallographic control of fracturing is lost.

PLANES OF LIQUID INCLUSIONS AND HYPOTHESES OF QUARTZ ORIENTATION IN DEFORMED ROCKS

Fairbairn (1942) has critically discussed three hypotheses of quartz orientation in tectonites. They are based on twinning, translation, and fracture. The twinning hypothesis is considered least likely to apply because little direct evidence has been obtained to substantiate such a mechanism. The translation hypothesis is also considered unproved because of the lack of direct experimental evidence. Deformation lamellae may originate by translation-gliding, but the lack of control (Ingerson and Tuttle, 1948) of their orientation by crystallographic directions in the quartz makes this hypothesis even less acceptable. The third hypothesis, based on the crystallographic control of fracturing, is considered most likely, and Fairbairn concluded (1939, p. 1490): "All the known quartz maxima in tectonites may be correlated with the fracture hypothesis . . ."; however, "the common oblique and *ac*-girdles are not adequately explained by any of the three hypotheses."

The fracture hypothesis requires quartz to break into needle-like slivers which are parallel to prominent crystallographic directions. The bounding faces of the needles must be crystallographic planes. The hypothesis is based on experimental deformation of quartz by Griggs and Bell (1938), who found that quartz broke into needle-like fragments bounded by prominent crystallographic planes when ruptured at elevated temperatures (450° C.) in the presence of sodium carbonate solutions. Quartz maxima are theoretically produced by rotation of the needles into shear planes with the prominent bounding plane parallel to *ab*.

It should be noted that the tendency of quartz to break into needle-like cleavage fragments was not encountered when the quartz was ruptured "dry" at room temperature with or without confining pressures. Thus the presence of solutions and the temperature may be important environmental factors controlling the behavior of quartz in tectonites.

The fracture hypothesis does not explain the ubiquitous girdles found in tectonites, and Billings (1942, p. 352) criticizes the hypothesis because ". . . we should expect to find many rocks containing such slivers of quartz. Actually, they are very rare." The evidence presented here for the orientation of planes of liquid inclusions in quartz indicates that quartz fractures much like an isotropic material under the conditions prevailing during the development of the planes and fails to give evidence for crystallographically controlled fractures in quartz. This fracture isotropism is further emphasized by Sander's (1930) investigation of four different types of fractures in quartz. *All four types show no lattice control of their orientation.*

Thus the geological evidence of the isotropic nature of quartz failure appears to render the fracture hypothesis as untenable as the twinning and translation hypotheses. Some type of growth hypothesis may afford a more plausible explanation of quartz orientation where the environment does not permit lattice control of the failure of quartz. Such hypotheses have been proposed by several workers. Griggs and Bell (1938, p. 1745), for example, suggest that differential solubility and the effect of the stress field on recrystallization might be operative to preserve certain orientations and obliterate others. Cloos (1946, p. 39) has suggested growth to explain quartz orientations in primary igneous and metamorphic rocks. As quartz is the latest mineral to crystallize, its orientation may be controlled by the pre-existing mica, feldspar, and hornblende. Cloos writes: "Growth in a girdle plane may be in any rhombohedron or prism—a large array of possibilities, the only restriction being that the availability of many other minerals in a definite direction exerts an influence in the growth and orientation of quartz."

In conclusion, no completely satisfactory hypothesis of quartz orientation is known. Until quartz fabrics are synthesized and the orientation processes are better investigated, this state of affairs will probably continue.

PLANES OF LIQUID INCLUSIONS AS EVIDENCE FOR OPENINGS FOR HYDROTHERMAL SOLUTIONS

Examination of a thin section of one granite (locality unknown) shows two sets of planes of liquid inclusions that differ in the character of the liquid. One set contains inclusions consisting of three phases—vapor, water, and liquid carbon dioxide—whereas the other set apparently contains only a vapor bubble and water. It is obvious that the two sets are of different age and that the character of the liquid available to fill the fractures changed with time. It is difficult to see how the composition of the liquid available to fill the fractures could change without introduction of material; therefore, a hydrothermal source for one or both of the liquids is suggested. This introduces the possibility that the fractures may provide avenues for migration of hydrothermal solutions.

At first thought it would appear ridiculous to postulate that such minute fractures could provide channelways for any considerable amount of solution; however, when a specimen such as no. 33 is examined and as many as two hundred planes are crossed in a traverse of only 1 cm., it seems that the openings provided are worthy of consideration by geologists studying ore deposits. It should perhaps be emphasized here that no megascopic evidence has been discovered that will indicate the presence or orientation of the planes. Also evidence of movement along the fractures has not been observed, although slight displacements would be difficult to establish.

SUMMARY

Planes of liquid inclusions are common in the quartz of igneous and metamorphic rocks. The planes consist of numerous individual inclusions, which are commonly planar parallel to the plane. The liquid inclusions are believed to have originated as fractures in the quartz grains, which subsequently became filled with solution. The solution-filled fracture is believed to change to numerous individual inclusions by a process of solution and deposition of silica. The process may have originated by a mechanism similar to that discussed by Riecke (1895), which requires that solution take place in a stressed crystal in the area under greatest stress and precipitate in the area of least stress. Temperature oscillation would also be a contributing factor, as would the fact that an irregular inclusion does not represent the lowest possible energy level; hence the process of solution and deposition will continue until euhedral cavities are formed.

This constant change in the character of the planes of liquid inclusions suggests a method of estimating the relative ages of two or more sets of planes. "Young" planes are characterized by their sheet-like inclusions, whereas "old" planes are composed of inclusions that are more nearly equant and are bounded, in part, by crystal planes. The relative age of two or more sets could be determined also by many of the methods used to establish the relative age between two or more sets of faults or joints.

Statistical studies of planes of liquid inclusions indicate that there is no control of the orientation of the planes by crystallographic directions in the quartz, suggesting that the quartz behaved like an isotropic material. Some evidence was found suggesting that the planes may be developed in the shear or tensional direction of the deforming stress pattern.

Statistical representation of the poles of planes of inclusions on a Schmidt equal-area net gives diagrams that have maxima up to 45 per cent. Maxima of 20 per cent are very common. Strong single maxima are most common, but split maxima have been found repeatedly. In the Washington, D.C., area two or more distinct maxima occur in the quartz veins but not in the gneisses. In other areas two or more maxima are apparently common. Many fabric diagrams of planes of liquid inclusions have incomplete girdles that are believed to be oriented about the *b* axes. Occasionally complete girdles are found.

Planes of liquid inclusions have a remarkably uniform orientation throughout the Washington, D.C., area, striking northwest and dipping steeply to the northeast throughout an area of 15 × 20 miles. Primary structures, such as contacts, foliation, and banding, strike predominantly north and dip steeply to the west. This nonsymmetrical relation is believed to indicate that the planes of inclusions are younger than the primary structures and therefore are not genetically related. No macroscopic structures could be found that were genetically related to them, although it is possible that some of the quartz veins are so related. The quartz orientation is, in general, believed to be genetically related to the primary foliation and lineation.

Detailed study of small drag folds indicates that planes of inclusions may develop in the shear direction of the deformation but that one set of planes parallel to one shear direction predominates.

Quartz deformation lamellae are present in the small drag folds. They also occur in the shear direction and are believed to be genetically related to the planes of liquid inclusions in the same folds. As lamellae are susceptible to recrystallization, it is not surprising that they are not commonly found with planes of liquid inclusions.

One investigator has described planes of inclusions that are believed to be related to crystallographic directions in quartz. However, the material studied may represent deformation at relatively shallow depths and at low temperatures. If this is the correct interpretation, then planes of liquid inclusions that are not related to crystallographic directions may represent an environment characterized by high hydrostatic pressure and probably also high temperatures.

A study of planes of liquid inclusions may throw some light on the mechanism of quartz orientation in tectonites. It has been established that quartz fails like an isotropic material in certain environments and that those theories of quartz orientation based on crystal structure control of failure are not applicable in such environments. Orientation by some growth process would appear to be more plausible.

Quartz may fail by shear or tensional rupture and in certain cases by deformation, which produces lamellae that may be a type of translation gliding. Failure by rupture along crystallographic directions has been noted by one observer and has been produced experimentally in the laboratory. These different phenomena are believed to represent failure under varying environmental conditions, and the optimum development of any one type of failure depends on such factors as confining pressure, temperature, composition, and the presence (and character) of solutions.

As many as two hundred planes of liquid inclusions have been found in a 1-cm. traverse of a thin section of a quartzite. Although the openings are minute, their abundance suggests that they are worthy of consideration as a channelway for the movement of gaseous and other solutions.

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Finally, I should like to express my gratitude to my wife for assistance in contouring fabric diagrams.

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ON THE GRANITE PROBLEM[1]

R. PERRIN[2] AND M. ROUBAULT[3]

ABSTRACT

Anomalies in the order of crystallization, the inadequacy of hypothetical aqueous residual and "pore" liquids, the complexities and inconsistencies of the theories of differentiation, the insuperable problem of granite emplacement (the "space problem"), the problem of the gneissic zones and the gradation of gneiss into granite, are all taken as evidence that the theory that granite is a product of crystallization of a magma is inadequate.

Evidence that granite originated by diffusion and reaction in the solid state is presented. Supporting this concept are the observed feldspathization of inclusions and country rock, double and reciprocal inclusions, digestion of transgressive dikes by the host rock, failure of "stoped" blocks to settle, microscopic evidence of replacement and corrosion, and the incomplete or arrested state of alteration of inclusions.

The eruptive character of granite dikes is an anthropomorphic or subjective impression. Conclusive evidence that some dikes are replacement phenomena is presented.

Though some authors admit granite to be polygenetic, it seems more probable that such a singular rock should be of one mode of origin, namely, diffusion and metasomatism in the solid state.

INTRODUCTION

The granite problem is again in the limelight. For many years the magmatic origin of granite was accepted without dissent; but now this dogma is being questioned by "metamorphists." There is vigorous defense by the "magmatists" against what Niggli (1942, p. 17) has called a revival of the "Universal Theory" without any new evidence or argument in support of it. In the past few years several authors have considered this problem. We appear to have been the first (1937, 1939), followed by Niggli (1942), Reinhard (1943), Read (1943–1944), and Raguin (1946). In the latest edition of *Eruptive Rocks*, Shand (1947) also presents a number of arguments for the classical theoriès.[4]

In this discussion we seem to represent the most liberal element, with Read following closely, whereas Reinhard, Raguin, and especially Niggli, support the more conservative views. Most petrographers in their writings have chosen one or the other theory. Some, on the other hand, (as, for example, Reinhard), have admitted that granite may be polygenetic.

The discussion has become more and more heated, and occasionally personal. P. Niggli (1942, p. 15) is particularly severe with R. W. van Bemmelen, who was the first to give full support to our ideas. Niggli makes no reference to our publications and therefore does not directly discuss our arguments. Nevertheless, he asserts that reactions in the solid state obtained in the laboratory are of little importance. He asks how the "emanations" were observed in the field and under the microscope, and why they rise. Finally, he asks what mysterious force tends to transform the sediments into a final product which is always the same, granite or granodiorite.

In a memoir which is itself not free of

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[4] Just after this article was completed, we received Memoir 28 of the Geological Society of America, "Origin of Granite" (conference at meeting of the Geological Society of America held in Ottawa, Canada, December 30, 1947). Our comments on this important work are appended at the close of this paper (p. 376).

contradictions, Max Reinhard (1943, p. 34) is clearly disdainful of "a few unpleasing phenomena" (*einige unerfreuliche Erscheinungen*). He concludes, after praising Niggli: "Je geringeres Gewicht man ihnen beimist, desto weniger werden sie die notwendigen Auseinandersetzungen über das Problem der Granitbildung belasten." On the other hand, Shand, though a "magmatist," does not agree with him, writing us recently: "Your contribution to the study is a most important one, and it will teach all of us to observe more closely and think more clearly." More recently Barth (1939; 1947, p. 181) cited our works in terms whose courtesy we thoroughly appreciate.

The importance of reactions in the solid state and their role in granitization are viewed in completely contrary ways. However, these contradictory judgments so positively expressed do not lessen the general impression derived from reading the works of contemporary petrographers, that most of them are obviously disturbed by the facts disclosed and the deductions drawn from them. However, many are evidently still influenced by the "magmatist" ideas that were taught them or which they themselves are teaching.

Diffusion and reactions in the solid state are unfamiliar to most petrographers. On the other hand, the prevailing concepts of magmatic differentiation appear consistent with data presented in the solidification diagrams. The classical hypothesis about the influence of water, "pneumatolysis," etc., taught as a synthetic, apparently coherent, construction, looks like a scientific certainty. For this reason it is necessary to review the whole question and to outline, very briefly, the hypotheses on which the magmatic theories rest.

THE "MAGMATIST" HYPOTHESES

SEQUENCES AND ORDERS OF CRYSTALLIZATION

Rosenbusch has stated what is still considered the normal order of crystallization of elements of eruptive rocks and, consequently, of granite. According to him, the first to crystallize and therefore the first to be deposited in the theory of solidification of granitic magmas are zircon, apatite, rutile, and sphene. All these are uncommon constituents of rocks and therefore of the original mixture. The early crystallization of rare constituents is inconsistent with what is known of melts in general: rare constituents crystallize from a solution only if there is a break in miscibility. However, zircon (zirconium silicate) and rutile (titanium oxide) are quite miscible in silicate melts and can be used in large quantities in metallurgical slags (whose nature is like that of rocks). This fact, in itself, is rather disturbing

After the so-called "accessory" minerals just referred to, there should appear orthosilicates with biotite, metasilicates, calcoalkalic feldspars, and, lastly, silica. Therefore, the deposition of potash occurs first during the formation of biotite, is interrupted by crystallization of the plagioclases, and is resumed during the crystallization of orthoclase or microcline. Such an interruption of deposition of an element in this way is contrary to what is known about crystallizing solutions.

A much more striking example of the interruption in deposition of a substance is that of water. During the crystallization of the magma, water appears as a constituent in early-formed minerals—biotite and, in many cases, hornblende, It is absent from the plagioclases, orthoclase, and quartz but reappears, at least theoretically, with the residual solutions (*rest Lösungen*) of Niggli. However, this water is rarely found in any mineral formed from such solutions. An interruption of deposition of this type is abnormal and even improbable. S. J. Shand has noted this anomaly (1944).

In a granitic magma (one with an excess of silica) it is not the silica that is thought to be deposited first. However, in the known silicate melts with an excess of silica, the silica is deposited first. How does one explain this anomaly? The usual concept is that water is in some way responsible. Assuming this and granting even the most improbable singularities, observation shows that, even from magmas of almost identical compositions, the order of crystallization may not be the same. We have cited (1939, p. 60) some of the numerous exceptions to Rosenbusch's rule. Shand, in the third edition of *Eruptive Rocks*, wrote:

> The only conclusion that we can draw from the foregoing discussion is that there is no such thing as a constant order of crystallization, applicable to all magmas or even to one magma under different physical conditions. The Rosenbusch rule has no theoretical or experimental justification, and even on the basis of observation it cannot be upheld. Yet the strange fact remains that the rule seems to be applicable to a large number of rocks. How can this be?

What hypothesis, then, can we formulate which will reconcile with physical laws of solidification these two numerous exceptions to the rule? Must we consider, as Niggli did, some mysterious "automorphoses" by the residual solutions of the magma? It would fail to explain the deposition of hornblende or olivine last in granites, a fact which led Sederholm (1923–1926) to the hypothesis of an ultra-basic residual magma, which is quite out of harmony with the general rule.

The formation of zoned feldspars with a sodic center or with alternating zones is not the least of the peculiarities of the crystallization of magmas. Several authors, including Phemister, Hil's, and Hibsch (Perrin and Roubault, 1939, pp. 63–65) have attempted to explain it. Norman L. Bowen, author of the anorthite-albite diagram, postulated a veritable dance of the feldspars in incomplete decantation in zones of magma with different temperatures (1928, p. 275). However, such an imaginative and unfounded hypothesis is inapplicable in the observed case of so-called "inverse zonation" (labradorite in the center and anorthite at the periphery), which is in absolute contradiction to the experimental diagram established by Bowen himself.

THE ROLE OF WATER AND RESIDUAL SOLUTIONS

Leaving out of consideration the water of quartz inclusions or of minerals such as mica, it is generally believed that water (or at least hydrogen) is necessary for the formation of granite (Perrin and Roubault, 1939, pp. 123–124). A considerable quantity is needed to explain the relatively low temperature of granite formation (in agreement with Goranson's experimental results). This water must disappear without producing the slightest distortion of the rocks by volume changes.

Granite must have lost its water, despite the very high pressures and relative impermeability of the rocks overlying the magmatic reservoir. On the other hand, lavas of the same composition arrive at the surface with 8 per cent water, which they retain, so that they solidify practically "saturated" with water (if Goranson's experiments are a valid guide to the solubility of water in lava).

Water plays a no less mysterious part

in the explanation of the "palingenesis" of solidified granites, i.e., those transformed on the spot, according to Sederholm, into new granites. These granites, having solidified and therefore lost their water, must now reabsorb water to account for their remelting at a necessarily low temperature. Every palingenetic stage requires the entrance and loss of water, again without leaving any evidence of deformation

The introduction of water quickly followed by its disappearance, which is needed to explain the lowered melting point of granite, is even more necessary to account for "residual" pegmatites, which, according to theory, were liquid at still lower temperatures What light do Goranson's experiments with alkaline quartz-feldspar mixtures shed on this question? Pegmatites should present fewer difficulties than the fusion of granite because both temperature and pressure were much lower. J. G. Koenigsberger (1942, p. 89) has given us the answer: Goranson's experiments have failed to prove anything in that respect. What happens, then, to the theory of pneumatolitic sequences and the older hypothesis?

But let us return to the subject of the residual liquids. These solutions are regarded as powerful agents of impregnation and transformation of the enclosing rocks (Niggli, 1942, p. 34). How, then, account for the extreme irregularity of metamorphism about granite bodies? In cases where it is almost nonexistent (Roubault, 1935, p. 21), must we assume the absence of pneumatolitic and hydrothermal stages?

These same residual fluids with their great impregnating power also have an equally important dissolving power over previously deposited minerals (Niggli, 1942, p. 21). This hypothesis championed by Niggli has been rightly attacked by several authors, including Bowen and Barth, who have shown that a residual liquid cannot dissolve previously formed crystals—it can only transform them into crystals with which it is in equilibrium.

It is an observed fact that pegmatite veins take on forms implying "corrosion" of enclosing rocks containing hornblende or biotite (early-formed crystals which ought not to dissolve), which makes it necessary to reaffirm the dissolving power of the liquid. Niggli invokes the "reaction principle" (1942, p. 21), which, however, does not permit a biotite or a muscovite granite to dissolve pyroxene or hornblende because of the antecedent position of these minerals in the reaction series. Such granite is "saturated" with respect to these minerals. Production of such crystals by the assimilation of sedimentary rocks is also not permissible. Eskola (1939, p. 379) writes that, in the light of Bowen's reaction principle, the granitization of limestone is generally rather unlikely. Yet the transformation of limestones by granitic endomorphism into real diorites is a widely known phenomenon, as Lacroix (1899) showed in his classic memoir on the granite of the Massif de Querigut in the Pyrenees. What new hypothesis should be formulated to explain this granitic endomorphism?

DIFFERENTIATION

The only scientific and logical theory published to explain the genesis of the so-called "differentiated" rocks from magmas is that of Bowen (1928). It is based on gravitative settling of heavy minerals, especially ferromagnesian, during the crystallization of the magma, and also upon a "squeezing-out" or filter-press action which separates the liquid parts from the already formed crystals.

Bowen's idea rests on an experimental basis. Niggli (1942, p. 71) believes that the criticisms of Bowen's theory are unjustified because they are concerned with the letter of the theory He amended it in his theory of the "complex gravitation crystallization differentiation" (*komplexe Gravitation Kristallization Differentiation*), according to which the first-formed crystals descend into warmer zones and redissolve, thus serving as vehicles for the transport of molecules. To this are added equalizing movements (*Ausgleichwanderungen*) and separations (*Abquetschungen*) by pressure. In like manner, according to Niggli, the assimilation is chiefly accomplished by the sinking of most of the blocks separated by "magmatic stoping." Niggli adds that we must explain not only how differentiation comes about but also how it leads to the formation of different, separated, partial magmas which are comparatively homogeneous in themselves. The mechanisms postulated by Niggli lead us to expect complete gradation between magmas and the rocks formed from them, but Niggli thinks that these mechanisms surely lead to relatively homogeneous partial magmas. He does not give reasons for this "surely," which, to us, seems hazardous. But in some so-called "differentiated systems," basic rocks overlie acid, which is contrary to theory and also to gravity. To interpret the relations on the basis of differentiation calls for new hypotheses involving squeezing, filter-pressing, or some other mechanism. In some places anorthosites overlie heavier gabbros (Nordingrå-Gebiet in Sweden; Barth, 1939, p. 95), which is to be expected under the differentiation theory, but elsewhere, as in the Adirondacks, the gabbros overlie anorthosites.

The formation of alkaline rocks (Barth, 1939, p. 91), the zoned rocks, such as the banded gabbros of Duluth, and the alternating zones of lherzolite and ariegite of the Pyrenees (Lacroix, 1899) offer special difficulties for the theories of magmatic differentiation. N. L. Bowen and Robert Balk have both offered hypotheses, but neither seems to us to be satisfactory.

At what point are we to stop? Let us repeat what Niggli has correctly said, that not only the differentiation but also the comparative homogeneity of the parts must be explained. The last case mentioned above is a good example of the problem to be solved by the "magmatists." Why not also attempt to explain the banded crystalline gneisses? Lest it be said that we exaggerate, let us point out that it is actually by differentiation followed by alpine dynamometamorphism that Casasopra (1939) explains the formation of certain gneisses of the Leventina Valley in Switzerland. These are identified by analysis as orthogneisses and also display very fine regular alternaation of thin gneissic zones of various compositions.

THE EMPLACEMENT OF GRANITE

Although many "magmatists" regard assimilation of already existing rocks as of great importance, Niggli, carried along by the inflexible logic of his theories (which allow granite very weak assimilative powers), is led to regard most granites as the products of juvenile magmas. However, being fully aware of the difficulty of introducing such magmas into already prepared openings (1942, p. 56), he seems to adopt Eskola's ideas of partial anatexis. In addition, he postulates subsidence of blocks detached by magmatic stoping. More and more authors assign a large role to assimilation, but Bowen himself acknowledges that granitic magmas are incapable of large-scale assimilation, and Raguin (1946, p. 194) says: "One would be tempted to invoke other, unknown, factors" ("On serait tenté d'invoquer d'autres facteurs inconnus").

The older work of the French school (Auguste Michel-Lévy, A. Lacroix, and Charles Barrois), together with that of Bucking and of more recent authors, such as Holmes and Reynolds (1947), have shown that in many cases the *mise en place* of so-called "eruptive granites" has taken place *à l'emporte pièces* without disturbing the sedimentary strata. In some cases traces of these strata have been found in exactly the position indicated by extension of the beds, either in the original form or as endomorphic modifications of the granite. What hypothesis will account for these relicts remaining in exact alignment within what was an eruptive liquid granite? These observations are of fundamental importance and are inconsistent with the "magmatists' " theory. Besides, the endomorphic parts of the granite are in many cases diorites or still more basic rocks. Theoretically, these rocks belonging to the eruptive rock facies went through a liquid stage, although, being anhydrous in the theory of differentiation, their melting point is much higher than that of watery granite. What new hypothesis must we create to explain this complete contradiction?

Another important point to which we shall refer later but which should be kept in mind has to do with the so-called "stoped" blocks. The theory of crystallization differentiation requires that crystals of chrysolite, pyroxene, hornblende, etc., the first to be formed in the solidification process, be decanted by gravity. However, Stokes's law demands that, with equal density and similar shapes, the rate of sinking should increase even faster than the increase in size of bodies. To explain the sinking of small crystals in a case where large or even great bodies located far within a granite mass (as in the Bergell area) have not descended may require another hypothesis. Only if the granite had a practically infinite viscosity—the viscosity of solids—can these observations be explained.

Related to the problem of granite emplacement is the emplacement of ultrabasic rocks: dunites, peridotites, and lherzolites, whose melting temperature at atmospheric pressure is very high (1,600°-1,800° C.). Here, too, hypotheses have been offered. However, Barth, who considered the matter at great length, unhesitatingly concluded that the problem is still far from solved (1939, p. 96).

ZONES OF ANATEXIS[5]

The emplacement phenomena briefly discussed above are more especially related to "eruptive" granites which clearly cut across country rock. It appears desirable to summarize some "magmatist" theories arising from the study of zones of anatexis.

It is not easy to understand the strict "magmatist's" theories of origin of the zones subjected to the action of Sederholm's ichor. The theories appear, in essence, to depend on the action of "residual pegmatitic solutions" capable of passing through all rocks, plus partial refusion—Holmquist's ultra-metamorphism and Eskola's partial anatexis imply such a selective fusion of the granitic elements, always with the aid of water. These theories are contrary to Wegmann's theory of transformations without fusion, by means of intergranular films. Eskola, however, seems to consider migmatization a sort of metasomatosis, but with the magma itself responsible

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[5] Including the "metatekt" zones of Scheumann.

("das **Magma** selbst als Träger fungiert") (1939, p. 378). This refusion, thus admitted, leads to a new hypothesis: Most of the rocks may have gone through a molten stage without disturbance of the sedimentary laminations and the folding of the original layers (outlined mainly by residual micas). (It is these observations which led Wegmann—quite rightly in our opinion—to reject the presence of magmas in these zones.)

It is common knowledge that rocks which have always been considered magmatic (such as pegmatites or aplites) have been observed in "metatekt" zones, or even in higher metamorphic facies. They commonly occur as isolated masses, clouds, *trainées diffuses* ("schlieren"), veins, or veinlike bodies, without being visibly related to a deep-seated mass and without any possibility of injection. It is difficult to explain that isolation of rocks which supposedly formed from magmas at the end of "liquation." A new hypothesis—*Ausblütung*, or exudation—came from the ideas of ultra-metamorphism. Niggli (1942, p. 33), after quoting from the words of Holmquist, Scheumann, and others, imagines a "molecular dispersion phase" (*molekulardispersen Phase*), whose migration would mean a partial remelting of the rocks, accompanied by concentration of some of the constituents ("Das Einspritzen ist oft nur ein Sammeln und ein Wandern"). It is also noteworthy that Koenigsberger (1942, p. 89), who collaborated with Niggli, considered the formation of anatexic pegmatites from exudations impossible.

It is necessary to devise still another hypothesis to explain already formed cavities of extraordinary shapes in which the "exudation" products crystallized. Many of these are filled with feldspars only, and, as the feldspars are anhydrous and therefore have a high melting point, an explanation should be (but is not) forthcoming. In addition, what explanation is there of clays containing 5 per cent alkalies (Roubault, 1935, p. 253) found in contact with granite but without exudates? Likewise, how explain masses, veins, "schlieren," etc., filled with soda-lime feldspars and even ferromagnesian minerals (hornblende, micas, etc.) which are theoretically crystallized first and are not a part of the "light, liquid, portion" (*leicht flüssiger Rest und Teile*), gabbro pegmatite observed by A. Lacroix au Pallet, or pegmatites of nepheline syenite reported by Barth (1939, p. 97)? Indeed, the composition of such rocks bears no relation to theoretical "residual" solutions, Therefore, generally speaking, the classical theories concerning the water of granites of the "anatexic zones" and of pegmatites leads to a law: The less hydrated minerals a rock contains, the more water it has contained. This really demonstrates objectivity in adapting theories to observations!

CONTINUOUS GRADATION OF GRANITE TO GNEISS

This is, again, an old subject of discussion. H. H. Read (1944, p. 51), after noting Élie de Beaumont's opinion that the origins of granites, gneisses, and mica schists are related, adds (p. 52): "This profound statement has been forgotten by many a thoroughgoing magmatist. For example, Niggli [1942, p. 7] believes that the association of granite and gneiss *complicates* the granite problem. For some of us, on the other hand, this association may clarify the problem." Niggli says: "It is obvious that these passages can be interpreted in two ways: the proximity of the granitic magma has created favorable conditions for an 'ultrametamorphosis,' or the magma is but the end product of the 'ultrametamorphosis' "

("Es ist selbstverständlich, dass sich diese Übergänge zweifach deuten lassen: Die Nähe des granitischen Magmas schuf die für eine Ultrametamorphose günstigen Bedingungen order das granitische Magma ist nur das Endresultat der Ultrametamorphose").

The hypothesis of convergence of structure and of composition between solid rocks which have been metamorphosed (even if they have been partially melted) and rocks which have crystallized from a melted magma remains to be considered. Niggli makes no reference to the zones of completely chaotic mixtures of granites and gneisses of similar composition which are well known to petrographers. How can one conceive of some of them having been completely melted while others were not, when the points were so close that the temperature was necessarily the same?

Evidence has apparently recently compelled Niggli (1947, p. 153) to make some concessions to the "metamorphists" by distinguishing not only "magmagranites" and "ultra-metagranites" but also "metagranites" resulting from metamorphism without any fusion whatever. He thus presents the hypothesis of convergence.

In short, the "magmatists" have gradually built up a whole series of hypotheses (we have listed over twenty, and there are many others) which they require to explain the facts. But when a concept needs such a concourse of hypotheses to support it (many of which are in opposition to one another), we can properly conclude that, however pleasant the structure may be to the eye, the foundation is weak.

Moreover, as we have seen above, this accumulation of hypotheses leads the "magmatists" to postulate several different geneses for granite, the choice depending on the requirements of the case. Reinhard, like Niggli, admits of three—a priori, this fact alone should have brought doubts to the minds of these authors, for granite is so special a rock that it seems unreasonable even to consider the possibility of convergence. It is difficult for us to imagine that such a crystallization did not in every case suppose an environment of the same nature: solid *or* liquid—one or the other, but not both.[6]

No doubt we shall be told that we also venture a number of hypotheses, perhaps still weaker, in our theory of granitization in the solid state. Nevertheless, in the following section we shall endeavor to show that our theories are based on indisputable observational data.

EVIDENCE OF ORIGIN OF GRANITE BY DIFFUSION AND REACTIONS IN THE SOLID STATE

GENERAL REMARKS

The belief in magmatic granite began with Hutton's (the father of the plutonian school) observations of the eruptive aspect of some granites, their homogeneity, and their apophyses and veins. The differentiation-crystallization theory added in recent years only supports the conclusions derived from Hutton's original observations.

Thus the belief in melted granite arises, first of all, from a subjective impression. "Anthropomorphic" was the term used by Maillet and Pavans de Ceccaty (1937) to characterize a similar attitude in tectonics. These authors wrote further: "Tectonics is not the only discipline in which the sensory imagination should be neglected" ("La tectonique n'est pas la seule discipline où l'imagination sensorielle doit être négligée"), and they showed the importance of logical reasoning, "the only instrument which can give access to these forbidden domains" ("le seul instrument qui puisse nous ouvrir l'accès de ces domaines interdits").

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[6] "Solid" is extended to include rocks previously discharged (*épanchées*) in the liquid state but solidified and, at a later period, affected by granitization.

In tectonics the "anthropomorphic" necessity of a plastic state of rocks no longer exists, and many authors admit the "flowing" of mountains by the action of gravity. A similar mental attitude seems indispensable when considering the petrography of granular rocks; for the problem is to determine whether a subjective impression is to prevail over rational deductions drawn from indisputable observations. With this attitude of mind, let us examine the fundamental facts which justify our way of thinking.

NIGGLI'S OBJECTIONS

First, we think it is necessary to answer briefly Niggli's three objections, mentioned at the beginning of this article. The first concerns the difficulty of reactions in the solid state. Our present state of knowledge of chemistry makes a discussion of this point superfluous. Our earlier memoirs (1937, 1939) contained an extensive bibliography on the subject. Since then Jens A. W. Bugge (1946) and Hans Ramberg (1945), in particular, have shown anew the fundamental importance of these reactions. No petrographer should deny, therefore, that diffusion of ions in crystalline lattices is an experimentally proved fact. Industry daily makes use of such controlled reactions, as in the preparation of magnesium by the action *in vacuo* of silicon on dolomite.

Niggli asks whether emanations have been observed in the field and under the microscope and why they rise. We could answer that they were observed as well as the *molekulardispersen Phase* of Niggli and that they rise for the same reason. Professor Niggli apparently believes that it is normal for the pegmatitic solutions resulting from differentiation to rise (although he does not say how), but he asks, if the granite is itself the result of deeper emanations, where do they come from?

In reality, observation shows that the phenomenon is not restricted to a rising of elements. Close study of the borders of a granite body, with or without zones of "eruptive breccia" or of "magmatic stoping," and of the evolution of the inclusions (*enclaves*) in any granite shows that certain elements came in while others went out.

Referring to such zones, Niggli wrote (1942, p. 41) that it is difficult to decipher from the final product all the stages of formation, because what has been a successive series appears in juxtaposition (*das Nacheinander wie ein Nebeneinander erscheint*). We do not agree with him; anyone who has examined border zones knows that it is possible to find practically all stages of transformation in the different inclusions, from the original rock (completely or nearly intact) to rocks hardly distinguishable from granite. This was observed by Lacroix in the Pyrenees (1898–1900). Raguin writes, about a schist-granite contact, that one passes into the other by a series of steps ("du schiste au granite, il y a ainsi continuité par une série d'échantillons"), and he describes examples of intermediate zones between the enclosing rock and the granite.

Simple visual examination of the inclusions obviously indicates that this gradation usually is characterized by increase in the proportion of alkaline feldspars and a decrease of the ferromagnesian elements.

We have called attention to the importance of diffusions *à double sens* (1946*a*), and recently Holmes and Reynolds (1947, p. 55) have given a new and magnificent example of this phenomenon. Such diffusions seem to be a general phenomenon, observed not only in cases of granitization but also in basic rocks. In the latter case nonalkaline elements, such as alumina or iron oxide, are diffused toward the enclosing rocks (Tilley, 1947, p. 104; Lacroix, 1899). The phenomenon also seems to be characteristic of metamorphism generally (Lapadu-Hargues, 1945; Perrin and Roubault, 1946). Therefore, the assertion that granitization is accompanied by diffusions *à double sens* is proved by verified facts. It is unimportant that the reasons for these deep exchanges are not yet apparent; the facts are there and are incompatible with the theories of simple influx of residual solutions, pore solutions, etc. On the contrary, they are observed in the case of reactions in the solid state.

Niggli asks what mysterious force always tends to transform the sediments into the same type of rock—granite or granodiorite. Niggli's idea is clear: it is normal for magmatic differentiation to tend toward a single rock type, but not for metamorphism to do so. He criticizes Van Bemmelen and Backlund (1942, p. 17), who think that the type "granite" is the expression of a tendency toward homogenization, toward a state of equilibrium. Let us note, first, that there is a tendency, not toward a single chemical composition, but rather toward an average chemical composition. This is made obvious by the very existence of Atlantic, Pacific, and Mediterranean provinces, by the diversity of composition of crystalline massifs—in brief, by the mass of evidence which has led to the theory of differentiation. In spite of this, the tendency toward homogenization remains an observed fact: granites tend to alkalinize their inclusions and gabbros to basify theirs, etc.

Recently, Lapadu-Hargues (1945) investigated the composition of a great number of "silica-alumina" rocks, from unaltered sediments to mica schists, gneiss, and even granite. He plotted the percentage of alkalines in molecular proportions as the ordinate and the percentage of alumina as the abscissa. For each rock type he drew a line enclosing the representative points of the alkali/alumina ratio, then connected the average points of the zones. He found, first of all, that the areas over which the points were distributed became smaller as the degree or metamorphism increased—in other words, the dispersion of analyses of rocks of the same type from different countries of the world decreases as the intensity of metamorphism increases. Moreover, he found that no break occurs in the line at the boundary of the gneiss and the granites. Thus the tendency to homogenization is evident in general metamorphism; with the transformist theory in mind it does not seem extraordinary that this tendency would be maximized in granite.

It should not be forgotten that geophysicists consider the heterogeneity of the earth's crust as only superficial. Consequently, it is not unreasonable to believe that sedimentary rocks, derived by alteration of the primitive crust, poorer than the crust in alkaline constituents, are in chemical equilibrium neither among themselves nor with their substratum. If at some time proper thermodynamic conditions arose, we do not see what would prevent homogenization by diffusion of ions, with the rising of alkalies from below.

We shall now discuss results of incomplete homogenization, namely, the so-called "eruptive" systems formed primarily from pre-existing detrital rocks. The world-wide abundance of granites and granodiorites testifies only to an ancient predominance of siliceous and silico-aluminous detrital rocks. This does not seem to us to be inconsistent with the existence of petrographic provinces. Let us now examine the principal facts which support our viewpoint.

THE FACTS

FELDSPATHIZATION OF THE COUNTRY ROCKS AND OF INCLUSIONS

Many authors have pointed out the existence outside of granite (either in the enclosing country rock, which indisputably remained solid, or in inclusions) of feldspars that are identical with those of the granite in every characteristic: composition, size, shape, color, inclusions, presence of microperthite, zonation (orthoclase surrounded by oligoclase, as in the rapakivi granite), etc. Large crystals of feldspar may even be found astride the granite contact. We have already dealt at length with this subject (1939, pp. 69 ff., pl. 1, fig. 15), taking into account the observations of other authors as well as our own. H. H. Read has independently cited similar examples. This extremely common observation is the basis of our ideas. Because forms of crystallization are so sensitive to chemical composition and environment, we concluded:

> The probability that crystals of feldspars, identical in every way, could have originated and grown under such radically different conditions as those of an acid liquid magma on the one hand, and of a solid rock (with a different composition, even mesocratic in some cases) on the other hand, is practically nil—and feldspars of a granite identical with those of the different, solid country rock, originated under the same conditions, that is to say, in the solid state.[7]

In a less precise, but more pungent wording, Read takes up the argument:

> One can, of course, believe that the two sets of feldspars, though admittedly identical, are really formed in two quite different environments, one magmatic and the other solid rock. I am tempted to remark, as the Duke of Wellington did on another occasion, that if one can believe this, one can believe anything! If this is really true, in view of the profound differences in the environments, I feel that a great part of the lore about magmatic crystallization becomes meaningless. I consider it more reasonable, and in the ultimate more honest, to hold that there is only one environment, that of the solid country rock [1944, p. 86].

Niggli and Reinhard made no reference to these observations or to the deductions drawn from them. Only Raguin attempted an explanation (1946, p. 184) by calling on Fenner's hypothesis, according to which, ". . . if three phases, solid, liquid, and gaseous, are in equilibrium, the deposition of crystals out of the liquid should be accompanied by the deposition of the same crystals out of the gases" ("si trois phases, solide, liquide, gazeuse sont en équilibre, le dépôt de cristaux à partir du liquide doit s'accompagner du dépôt des mêmes cristaux à partir de vapeurs"). However, the same author very prudently wrote also:

> If there is, in cases of this sort, an objection to the passage of the granite through the liquid state, one may answer that, among the prevailing chemico-physical conditions, no one knows which determine the composition and the facies of the feldspar crystals formed. It is possible that, among these factors, the liquid state, more or less advanced, is not essential. At

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[7] "La probabilité pour que des cristaux de feldspaths, identiques en tous points, aient pu naître et se développer dans des conditions aussi radicalement différentes que celles de cristallisation dans un magma fondu acide, d'une part, et de cristallisation dans un milieu solide, de composition différente, allant jusqu'à des roches mésocrates, d'autre part, est pratiquement nulle et les feldspaths du granite, identiques à ceux des terrains étrangers solides, ont pris naissance dans les mêmes conditions, c'est-à-dire dans le solide."

the moment of crystallization of the feldspars, the granite is no longer liquid, but mixed, with numerous solid particles in suspension. Reciprocally, the surroundings are soaked in the solutions. The physical states are perhaps no longer very different.[8]

Here again Wellington's remark comes to our mind. So, before changing our opinion, we shall await an incontrovertible explanation of the phenomena of feldspathization based on the magmatic theory.

THE PHENOMENON OF DOUBLE AND RECIPROCAL INCLUSIONS

The formation of crystals in the country rock or in inclusions identical with those of the granite is not restricted to the alkaline feldspars, particularly orthoclase. Plagioclase and ferromagnesian minerals may occur in a similar manner; the examination of certain contacts gives the impression that granitization grows somewhat like a mold which invades the enclosing rocks little by little. There are either numerous isolated crystals on the bottom of the old rock corresponding to what we have called the "progression de la granitisation en cristaux isolés" (1939, p. 75) or actual granitic areas inside the rock (1939, pl. 1, photo 3). The two phenomena are closely associated. Read (1944, pp. 86–87) has reported similar observations.

The continuity of the phenomenon is demonstrated by microscopic examination. In the case of the granite of Trebeurden, Bretagne, we have observed crystals of orthoclase with quartz, or crystals of orthoclase associated with plagioclases (1939, pl. 3, microphoto 4). Finally, there occur masses of granite existing as *inclusions within the inclusions of the same granite*—it is to this phenomenon that we have given the name of "double inclusion" (*double enclave*).[9]

Also observed is granite which contains inclusions of the surrounding rock, and, reciprocally, this rock contains inclusions of granite. This we term the "phenomenon of reciprocal inclusions." We have given an example of this from Bénodet, Bretagne (1947), and analogous observations have been reported by Wenk (1943, p. 267) and Barth (1947, p. 178).

Some "magmatists" will be tempted here to invoke *Ausblütungen*, which, according to the proponents of this hypothesis, are mixtures with minimum melting points, the melting points having been lowered by water, which afterward disappeared. We have already pointed out that a number of hypotheses are needed to support *Ausblütung*. In the present case an explanation by means of *Ausblütung* with minimum melting point is impossible by reason of the observed continuity from the individual feldspar crystals (with a high melting point) to whole granitic bands. An indisputable geologic fact remains: there have existed, in rocks in contact with granite or enclosed within it, favored points where there were formed crystals identical with those of the granite, followed by the formation of whole granitic zones.

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[8] "S'il y a, dans les faits de ce genre, une objection au passage du granite par l'état liquide, on peut répondre que, parmi les facteurs physico chimiques régnants, nul ne sait quels sont ceux qui déterminent la composition et le faciès des cristaux naissants de feldspaths. Il est possible que, parmi ces facteurs, l'état liquide plus ou moins avancé ne soit pas essentiel. Au moment de la cristallisation des feldspaths, le granite n'est plus liquide, mais 'mixte' avec de nombreuses particules en suspension. Réciproquement l'auréole est alors imbibée de microsolutions. Les états physiques ne sont peut-être plus très différents."

[9] Raguin (1946, p. 183) explains *double enclave* as the presence of separate crystals of orthoclase in inclusions of the granite; we consider that a simple case of feldspathization of the inclusions.

DIGESTION OF DIKES BY GRANITE

Dikes of several kinds, with sharp contacts and parallel edges (diabase, dolerite, aplite, granulite), intersect a granite, but in places some granite "apophyses" project into the dike or may even replace it entirely and retain a few widely separated and more or less transformed inclusions of the vein (Perrin and Roubault, 1939, pp. 127–129; 1947).

Sederholm (1923–1926) observed similar phenomena, which he ascribed to a "reviviscence" of the granite. The latter, after having been cut by the dike, would have remelted and been injected into some fissures opened in the dike. Raguin (1946, p. 19) reverts to this hypothesis in studying a dike which intersects an old granite, described by Sederholm. He adds: "A paradoxical fact, this granite is at the same time much older and much younger than this basalt" ("Fait paradoxal, ce granite est à la fois beaucoup plus vieux et beaucoup plus jeune que ce basalte"). It is simple and positive!

How did the granite melt again? How else except with the help of water because water is necessary for the refusion. The solidified granite all at once received a lot of water as the temperature rose; but, in spite of an implied large increase in volume, no mechanical strains resulted, as there is no evidence of deformation in the intact or unaltered portion of the dike. However, some fissures were opened, and the "revivified" granite was injected into these fractures; then the water went away, and all was normal again. The water must not have reached the dike or else (according to the classical theories) it would have been melted because its melting point ought to be lower than that of granite, even if it were aplite or granulite.

Unfortunately for the theory, in many cases there is not only the appearance of injection but actually of digestion—of magmatic stoping or complete digestion. In some examples even the edges of the dike remain unchanged in the digested parts (Perrin and Roubault, 1939, pl. III, fig. 3; 1947).

We shall not, for the time being, form a hypothesis concerning the genesis of the dikes; but, whatever the hypothesis, the contents of basalt or diabase dikes were injected in a liquid state. The related magma then intersected solid rock, and the granite was substituted in these necessarily solid rocks; the granite digested in places the earlier dikes as it digested the enclosing rock. This granitization was possible only in the solid state; for, in the midst of the melted granite, some of the dikes of aplite or granulite, theoretically more fusible, would have melted and been mixed into the granitic magma.

NONSETTLING OF INCLUSIONS

Inclusions, very common in contact zones, even some of great size and heavier than granite, have remained in place. All the theories of Bowen and Niggli are based on the settling of little crystals of olivine or amphibole; but these same authors do not find abnormal the failure of enormous inclusions to settle. Such an attitude leaves us amazed!

Some authors believe that the granite of border zones was extremely viscous. But we note that this granite, rich in pegmatitic solutions, cannot have high viscosity and, at the same time, the extreme fluidity necessary for the formation of lit-par-lit injections.

Observation shows that inclusions rich in "ferromagnesian" minerals are very common in the heart of granite masses. Some of them are *homoeogènes* and others *énallogènes*, according to Lacroix's terminology. For all of them the melting point is (according to the "magmatist" theories) higher than that of the granite, and their density is also greater. Although located in zones where there was no exceptional cooling, they did not settle.

Lacroix and many others have emphasized the extreme slowness of the granitization processes. The failure of the inclusions to sink is compatible, therefore, only with a state of "infinite" viscosity—the viscosity of solids.

EVOLUTION OF CRYSTALLIZATION

We have considered, first, some fundamental field observations and then some equally basic details of crystallization of the minerals of granite and of associated rocks. Lack of space prevents discussion here of the phenomenon of corrosion between crystals (Perrin and Roubault, 1939, p. 160, pl. 5, microphotos 7 and 8). Other authors, in particular Shand (1942) and Bugge (1946), have made similar observations of this very common phenomenon.

We shall cite as an example only the plagioclase "islets," unquestionably "corroded," surrounded by quartz and arranged in such a manner that the twinning planes are rigorously parallel. In addition, twinning bands of the same width are clearly aligned from one "islet" to another. In such a case dissolution is out of the question. The new crystals were substituted for the old ones by the substitution of the new lattice for the old, and this was possible only by the diffusion of ions of alumina, sodium, and calcium, which diffused through the crystalline lattice of the quartz to become fixed at more or less distant points. The same is true of hornblende replacement by orthoclase (Perrin and Roubault, 1939, pl. 3, photos 6 and 7); some calcium, iron, and magnesium ions diffused through the crystalline lattice of the orthoclase, and potassium ions diffused in the opposite direction to nourish the orthoclase lattice. Here is unquestionable evidence of diffusion in two directions in a solid state—diffusions observable on the microscope stage.

Water or, more correctly, hydroxyl OH, is but a special case of the diffusion of hydrogen ions. Bugge (1946) insists that, even if the constituents of a hydrated mineral were substituted in such a fashion, the reaction should be considered as taking place in the solid state.

Even if the processes had been "set off" by gases, pore solutions, or intergranular films, microscopic observations demonstrate that diffusion of ions through the crystalline lattice was necessary. As an example, Shand (1942, p. 417) described "some plates of hornblende which enclosed the corroded but undissolved remnants of the three minerals [hypersthene, augite, and plagioclase]"—how else may we explain the transformation of a pyroxene fragment wholly enclosed by the hornblende?

Along the same line we have never understood how authors who invoke gases or pore solutions can thus explain the formation, in a solid rock, of a metacryst of orthoclase which replaces an aggregate of quartz, biotite, and plagioclase; replacement requires that iron, magnesium, and calcium have gone out and potassium come in. In the light of modern physicochemical knowledge we see only one possible explanation: diffusion in two directions through the lattice.

RATE OF GRANITIZATION

Along the borders of many granitic systems (Mont-Blanc, Pyrenees, Massif de Bergell, etc.) are found inclusions of the country rocks. Some of these inclusions, located very far from the contacts, are at different stages of transformation. The very existence of incomplete transformation revealed by the study of these inclusions indicates extremely slow processes. These processes were certainly much slower than the known speed of homogenization in a liquid state. They were also much slower than the speed of solution of a silicate solid in a molten silicate bath. Let us recall here the perfect miscibility of silicates.

The slowness of reactions of diffusions in the solid state has been cited as an objection by some. To us it seems, on the contrary, indispensable to explain the facts, even in proved cases of diffusion over long distances. Taking into account the enormous duration of geologic time, experimentally determined speeds of reactions in the solid state would be much too great for the inclusions to survive.

When Bugge (1946, p. 48) writes: "The chief objections against such a long distance diffusion of ions through the crystals is the small rate of diffusion. But I do not think this is a deciding objection. The total development of a mountain range comprises a period 250–300 million years," we quite agree with him, but with a difference. This slowness is a favorable, rather than a negative, argument for our viewpoint.

THE ERUPTIVE CHARACTER OF GRANITE DIKES AND APOPHYSES[10]

The eruptive character, or "aspect," is the basis for the classical conception of melted granite; it is therefore normal that it should form the "anthropomorphic" objection to our conceptions. Following are some facts.

We have previously described (1939, p. 95) the occurrence in some crystalline granitic masses of crystals of feldspar or amphibole in what appear to be apophyses or veins with parallel edges, but of coarser grain than the crystalline massif itself. The injection of a liquid in a preexisting slit is out of the question. We have published photographs of this phenomenon (1939, pl. 1, photos 6 and 10) and have since observed new examples of the same thing on the Brittany coast (in the Argenton area, in Finistère).

We have also published observations on granite "veinlets" of the Anse de Sciottot, near Flamanville (1938*a*). Transverse sillimanite-rich zones in the granite dike appear to be but prolongations of such zones in the country rock, and they thus preclude the possibility of a liquid injection of the dike. A very fine and quite similar example has just been described by Bearth (1945). He figured a granite apophysis of the "Mont Rose" in clear discordance with the paragneiss and described some garnet beds of the gneiss which continue into the granite. He concludes that the granitization is thus the result of a metasomatic process or replacement.

No comment appears necessary. Injection of liquid is out of the question in this splendid example in which the garnet lines remained in place, just as zones of inclusions commonly do. We have in this example obvious proof of the emplacement of granite dikes by means of simple chemical exchange, to the exclusion of any injection.

The alignment of discrete and separated feldspar crystals, which we as well as Charles Barrois and Auguste Michel-Lévy have described (1939, pp. 96–100), is a related phenomenon. Did not Barrois use the words *filons moniliformes* to characterize these peculiar alignments? After our first memoir we published a new study, with photographs (1938*b*), on the phenomena of magmatic stoping at St-Jean-du-Doigt and also observed another splendid example at Argenton, Finistère.

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[10] Some authors (Wegmann, Green, Backlund, MacGregor and Wilson, and Reynolds) are quoted by Read in support of the hypothesis of "mobilization" to explain the eruptive aspect. We believe it unnecessary to envisage, for most cases, any major movement—in our opinion, and therefore in our analysis, the explanation of the eruptive aspect is different.

In all these cases, in order to explain such alignments by the hypothesis of liquid injection, we must imagine some mysterious fissures which would later heal over without leaving any evidence. Must we also, perhaps, admit of cavities prepared in anticipation of the injections? In the absence of such openings, what is the mysterious composition of a liquid which, by solution of hornblende and plagioclase, would lead to orthoclase? If we are not to search for an interpretation which would at any cost save the idea of liquid injection, this phenomenon must be considered, as being of the same nature as that responsible for feldspars in the rocks bordering granitic masses.

The feldspars found in the country rock are the crystal precursors of the normal granitization front; in like manner, the aligned feldspars are the precursor crystals of the apophyses or dikes of granite or pegmatite. *We must therefore accept the idea that, in some cases, granitization follows preferred directions; it results, therefore, in the construction of apophyses or dikes. The phenomenon is the same whether granite proper or pegmatite or aplite is in question.*[11]

ARE THERE DIFFERENT WAYS FOR THE GENESIS OF GRANITES AND ASSOCIATED GRANULAR ROCKS?

Taking the advice of Maillet and Pavans de Ceccaty (1937), we have tried to neglect *imagination sensorielle*, to leave out of account the *représentations concrètes* to which our senses are accustomed, and to outline a theory for the genesis of granites from the observation of incontestable facts. Have we the right to extend to all granites the conclusions which have forced themselves upon us?

Let us make it clear that we are concerned only with the process of granitization. We do not consider the fact that a granite can be formed by substitution from very different rocks—sedimentary, metamorphic, microlithic (such as lavas), or even other granites. Thus we do not exclude the possibility that some granites or some granular rocks could have been formed through mere recrystallization in a solid state of eruptive rocks. One of us (Roubault, 1935) has described aplite veins in Kabylie de Collo, where the petrographic relationship with the rhyolite veins of the same area is obvious. We think it quite possible that the recrystallization of veins whose contents were originally formed by a liquid lava has resulted in a granular rock of the same or very similar composition.

All the observations that we have made lead us only to the belief that some diffusion, however slight it may be, is necessary for crystallization into a true granite with coarse constituents. But the essential problem is to determine whether granitic texture always originates in the solid state. This is the crucial point.

An increasing number of geologists, at first incredulous, now admit the possibility of granitization in the solid state but are reluctant to admit its prevalence. They tend to hold the opinion of Read, who wrote: "There are granites and granites." This author does not deny the possibility of a magmatic origin for little granitic bodies whose emplacement is not an insoluble problem; on the other hand, he repudiates it for most granitic masses and especially the big ones.

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[11] After considerable delay there has just come to our attention an article by H. G. Backlund (1943), in which we are pleased to find ideas convergent with our own. Wegmann's excellent observations (1943) concerning progressive granitization of the Amistock conglomerate in Greenland also adds new evidence favoring our thesis.

As noted elsewhere, Max Reinhard, too, believes in the formation of granite without melting, by metasomatosis, but he also admits two other possible geneses: juvenile granites and palingenetic granites. Niggli (1942, p. 83) approvingly quotes Frank Grout's remarks:

> Recent writers have now suggested that igneous-looking rocks resulted from metasomatism in so many places that they assume the process is established, wholly fail to see the complexity of the problem, and quote each other as authorities, but not one area has been described convincingly. It seems that the enthusiastic advocates carry their idea too far, applying it to masses several miles across; the conservatives admit a few feet; the truth may lie between, but the conservatives have the weight of evidence on their side, and the burden of proof is on the advocates of extensive granitization.

Grout seems to have summarized the situation quite well.[12] The present conflict is epitomized by the following quotation from H. H. Read (1944, p. 86):

> The inability of the magmatists to interpret the plain field-facts arises from the inertia of a century of belief in the magmatic origin of all granitic rocks. The inertia has been increased for many by the addition of weighty physico-chemical arguments which "explained" the phenomena. It takes a certain amount of courage to go around to the other side of these problems and look at them from a completely different situation.

The weight of evidence claimed by Grout does not seem very heavy to Read. Having for thirteen years had the courage to differ from the authorities, we continue to behave as representatives of the "extreme left," not because we mischievously want to look like revolutionaries, but because we know of no fact which would justify our setting a boundary between the "few feet" and the "several miles"—and we do not want to say, out of caution, something that we do not believe.

At the risk of repeating ourselves, we think it necessary at this point to outline the granite problem as it appears from the writings of even the most "magmatist" authors. None of them (Niggli, Raguin, Grout, or Reinhard, for example) would think of denying that rocks indistinguishable from granites were formed in the solid state by metamorphism or metasomatism. Nor would they deny gradation from granite to alien rocks of sedimentary origin, in particular, crystalline schists, paragneiss, quartzite, etc. However, the "magmatists" consider that only border phenomena are in question; Grout's "few feet" are explained by the phenomenon (or, more correctly, the word) of "convergence," so they go on thinking of the interior as resulting from the crystallization of a liquid. Raguin (1946, p. 68) admits such a convergence between a granite with endomorphic alteration and its sides or its exomorphosed inclusions. The same convergence is envisaged even where the granite border has not been endomorphically altered. Grout, who does not deny the continuity, evokes the difficulty of proving metamorphic origin inside the massif. As to Reinhard, he admits the total structural convergence of metasomatic, juvenile, and palingenetic granites.

Behind all the words and reservations is the following important hypothesis: Similar granitic structures can result from processes as radically different as crystallization from a liquid and transformation of a solid rock by means of chemical changes in the solid state. Niggli (1942, p. 60) and Reinhard (1943, p. 24) say so plainly: "Material transfer may result in structures convergent with magmatic structures; even if the metamorphism takes place in the solid state," and "Granite may be formed from sandstones and clays through such metasomatic-metamorphic processes without liquefaction. These granites do not differ from magmatic granites either in mineral composition or in fabric" ("Stoffwanderungen können zu Konvergenzen mit Metamorphose im festen Zustand erfolgte," and "Es können aus Sandsteinen und Tonen durch solche metasomatisch-metamorphe Prozesse ohne Enischmelzung Granite gebildet werden, die sich weder in Mineralbestand noch im Gefüge von magmatisch entstandenen Graniten unterscheiden"). The thought behind the words is obvious: "endomorphe" supposes a melted state, "exomorphe" a solid state. We wonder if these scientists realize the full implications of the hypotheses thus expressed, numerous singularities of granite structure and composition being considered; for in the two cases the characteristics are the same down to the slightest detail. To put it emphatically: *the probability of producing such similar structures by such different processes seems to us absolutely nil;* and we think it wrong to postulate such fundamental differences in genesis when observation reveals only continuous gradation. Therefore, either all the observed rocks passed through the liquid state or they remained in the solid state. Nor have we the right, in our opinion, to imagine a semimelted state between the two.

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[12] In 1941 Grout knew only one of our granite papers—the one about "magmatic stoping" of St-Jean-du-Doigt, which he quotes; accordingly, he makes no reference to our other observations and arguments.

Before closing, we should like to point out that the observations which support our arguments were made by many authors independently in many parts of the world.

As several authors have noted and as Read has emphasized, the pavements, the monuments, and the building fronts of the cities of the world offer countless examples; it seems as though they would shout the fact to petrographers. We would not go so far as to wish, as Read humorously did (1944, p. 50) regarding the twelve columns of Wisdom, that characteristic chips might be detached by air bombardment. But we both have traveled extensively and made the same observations from northern Europe to central Africa and across the United States. As Read has written, "Every city has on free display magnificent examples of granitization and migmatization."

We are even astonished that the phenomenon which is at the base of our conceptions did not attract more attention among geologists. Unquestionably, we are concerned with a very general phenomenon, visible in all the granites of the world, which may therefore be considered characteristic of granite origin. Let those who want to be convinced of it follow Read's advice and look at the "free displays" of their towns.

It is the same with regard to the *double enclave* phenomenon and the digestion of veins. As to diffusion in two directions in the border zones and in the inclusions, we shall only refer to the work of H. G. Backlund and of Bugge on arendalites (1946). We have ourselves pointed out the very general occurrence of intercrystalline corrosion.

Under the circumstances, we really do not see why the importance of the observations should be limited to *some* granites and to distances of but a few feet, especially (as Read pointed out) because it is in the larger granitic systems that emplacement in the molten state is inconceivable.

We think also that it would be dishonest of us to imagine a different genesis for diorites, gabbros, and other granular rocks. The diorites and gabbros, chiefly the former, often constitute an "endomorphic" alteration of granite. Nor should we forget the digestion of country rock with "endomorphism" of gabbros (Lacroix, 1899), the basification of acid inclusions in gabbros, banded gabbros, and the alternate beds of ariegite and lherzolite.

At the risk of having Niggli accuse us of willingness to revive a "universal theory" incapable of explaining the multiplicity of geologic facts, we shall indicate briefly why we believe the mechanism of diffusion *à double sens* in the solid state has an importance far beyond the case of the so-called "igneous" rocks.

First, as we have already written (1939, p. 145), we think that some "extruded" rocks, i.e., lavas, underwent such diffusions after cooling, with the formation of phenocrysts, which were by no means formed in the first stage of consolidation. Later we observed at St-Jean-du-Doigt in Brittany feldspathization of some of the enclosing rock, leading to the formation of feldspar crystals, which gave these materials an absolutely clear appearance of two times of crystallization (1938*b*).

Observations on other lavas by other authors lead us to the belief that later diffusions have had a very clear influence upon the appearance of these lavas in the solid state.

Approaching, now, the problem of metamorphism in general, such work as Lapadu-Hargues's statistical study demonstrates fractional diffusion (*diffusions fractionées*) in two directions—a process identical with that of granitization. Furthermore, in what is apparently a case of pure dynamic metamorphism, at Tré la Tête in the Mont Blanc system, we showed (1946*b*) by analysis and thin section the existence of these diffusions. This confirmed our idea that there is no break (*coupure*) between regional and dynamic metamorphism.

Lastly, in many gneisses and mica schists, we find the same phenomena of superposition in crystallization between crystals and of crystal corrosion, as in granites.

In the tradition of the French petrographic school and of Auguste Michel-Lévy, we do not think it possible to prove that granitization, contact metamorphism, and general metamorphism are of different natures. Being perhaps less prudent than some other authors (like Bugge, who, however, emphasizes the important role of reactions in the solid state), we do not consider other mechanisms, like liquid infiltration, very important. We would except, of course, the cases of lavas and some metallic vein phenomena. As to Wegmann's intergranular films, they do not seem to have played a very important part.

The following reasons lead us to the above conclusions:

1. Crystal corrosion necessarily implies diffusions in the lattice, and consequently we see no necessity for recourse to another phenomenon.

2. Geological observations seem to show that the larger the crystals, the more intense the exchanges, although the opposite should be true if the major role were played by diffusion along the adjoining surfaces because these surfaces are smaller in proportion as crystal sizes increase.

Lastly, we are more and more convinced of the major role played by diffusion in the solid state in the genesis of many metalliferous ore deposits, particularly igneous deposits. In 1934 one of us had already expressed this idea as a hypothesis (Perrin). The many discussions now appearing in the literature concerning the nature of some deposits, termed "hypothermal," "contact," or "segregation," and the increasingly frequent observation in ores of replacement structures and of crystallization with corrosion show that their genesis certainly depended upon the same mechanism as that of granular rocks.

In closing this paper, Read's humorous sentence comes to mind (1944, p. 84): "Nobody except him who has learnt less and forgotten less than any Bourbon can deny the validity of the conversion 'of solid rocks to rocks of granitic character without passing through a magmatic stage.' "

We shall only point out to Read that it is not easy to get away from ideas which have impregnated one's mind since youth to such a degree that they have become a habit of thought, a habit of reasoning, a habit of language. Perhaps it is even harder not to be guided by subjective "anthropomorphic" impressions.

As for us, we are convinced that not without an inner qualm do some eminent scientists make use, for the lack of a better one, of the word "convergence."

SUPPLEMENTARY REMARKS

We had finished the preceding article when we received the report of the symposium on the origin of granite held at the December, 1947, meeting of the Geological Society of America. Perusal of this report leads us to add the comments below.

1. Professor Grout, citing Read's discussion of feldspars, writes: "This has no basis in chemistry or in logic." Professor Shand also sees no reason why the same feldspars cannot have their origin in the country rock and in the magma under the same pressure and temperature conditions. Both completely exclude the fundamental influence on crystal growth of the medium in which the crystals were formed. We repeat that there is no probability that crystals identical even to minute details (shape, color, inclusions) could be formed in part by replacement of solid crystals in a solid medium and in part by crystallization starting from a molten magma. In order to be sure, we have discussed the matter with eminent nongeological specialists, and their opinion exactly concurs with ours. Furthermore, we should like to learn how Grout and Shand explain the porphyritic crystals which lie across the boundary of the granite and the country rock, as well as granitic bodies entirely isolated in country rock near the contact.

2. Dr. Bowen criticizes but one citation out of a hundred concerning reactions in the solid state—Greig's work on silica bricks—thinking, perhaps, that this one would throw doubt on our entire theory. He makes no reference to Adams' observations or to Perrin's experiments on refractories, in the course of which not the slightest trace of liquid could be found. Our experience with metallurgical linings enables us to state that if we mentioned a "purely hypothetical intermediary liquid" (*intermédiaire liquide purement hypothétique*) to explain the reaction mechanism, it was because we thought that, in the case of such phenomena, the existence of liquids is not the cause but the result; diffusion of elements in the solid state lowers the melting points of refractories while altering their composition, and liquid parts are formed.

Bowen cites, as evidence against our theory, the existence of zoned feldspars, apparently claiming that the speed of homogenization of heterogeneous crystals of the same crystalline system are, in principle, the same as all rates of ionic diffusion, particularly those which lead to the replacement by substitution of one crystalline network for another. This is certainly erroneous (Perrin and Roubault, 1937, p. 58). Homogenization is a phenomenon which in many cases requires high temperatures. Bowen (1926) has shown that for the solid solution of NH_4Cl–NH_4NO_3 the homogenization temperature is very near the melting point. In our opinion it is much harder to understand why homogenization did not take place at the time of the certainly very slow cooling of a previously molten magma than is the case of reactions in the solid state, which experiments have shown to take place at temperatures much lower than the melting point (for example, BaO and SiO_2 at 357° [Hedwall] CaO+sillimanite, resulting in metasilicate of lime and alumina at 530° [Taylor], etc.).

Why did homogenization not take place in the process considered by Bowen, which requires much higher temperatures than ours? Everyone knows that speed of diffusion increases very rapidly as the temperature increases. Bowen would give much better support to his theories if he would indicate how they can account for plagioclase with completely reversed zoning: labradorite in the center and anorthite around it. How can they explain the existence of feldspars in diorites, theoretically magmatic in origin but which are, in fact, one of the endomorphic facies of granite (Lacroix, 1898–1900)? This phenomenon of inverse zoning is in direct contradiction with Bowen's own experimental diagram of solidification.

In a different way, Bowen criticizes the theories of Ramberg and Bugge, which are merely the first tentative efforts at the erection of a theoretical basis for diffusion in the solid state as applied to geological problems.

We believe the logical way to make progress in this problem must be (*a*) the making of field observations, (*b*) the drawing of scientific deductions from these observations, disregarding "anthropomorphism," and (*c*) the construction of a general theory of the processes.

It is the general aspect, as seen in the field, which has led us to believe in the process of diffusion in the solid state.

It is almost inconceivable that Ramberg and Bugge could have succeeded, at the very first attempt, in formulating theories which could explain everything in a practically unexplored field like that of large-scale ionic diffusion. These authors, nevertheless, deserve a great deal of credit, as does Bowen for his attempt to set up a scientific magmatic theory. With a few exceptions, we might have agreed with him, if accurate observation had not shown that the facts demanded concepts of an entirely different nature.

3. Professor Read, although he favors a large part of our theories (since 1942 he has accepted unaware practically all our essential arguments concerning feldspars in the country rock), calls us "enthusiasts." He considers that we go too far and that there are "granites and granites," although he seems to leave but a small place to magmatic granites.

We objected to the theory of total convergence of structure between rocks which would originate from entirely different genetic processes; we will apply to geology a principle applied in other fields, that is not to introduce an artificial hiatus where none has been observed.

We note that in the course of the very

interesting discussion at the Ottawa meeting, many petrographers asked for definite criteria for distinguishing the two kinds of granites, as there seemed to be no such criteria. This clearly shows that no one knows where the "hiatus" should be placed, and it is an encouragement for us to remain "enthusiasts." The only real, visible distinction, as indicated by Dr. Chapman, is the nonintrusive character at depth and the intrusive at higher levels. But it so happens that the same body of granite has the two aspects, depending on the part of the border under consideration. Besides, feldspathization of country rock is considered clearer and more conclusive evidence of the intrusive origin.

These remarks justify the special importance which we attach to evidence derived from accurately obtained samples in deciding that the intrusive aspect is not evidence of liquid flow but that, on the contrary, it coexisted with unquestionable cases of replacement in the solid state. We are pleased to note Goodspeed's striking example of granodiorite dikes with zoned feldspars, formed by replacement, and Nelson's similar observations. The simple anthropomorphic impression of injection of a liquid can no longer be taken as proof of such injection.

Participants in the symposium discussed the percentage of granites and granodiorites originating from magmas or by replacement in the solid state (despite convincing criteria for distinguishing them), and 15 per cent seems to have been suggested as a minimum for the latter by less optimistic members. In 1934 and 1935 Perrin came to the conclusion that metamorphism took place by diffusion in the solid state and that granites were formed as a result of this process. He would have been called an enthusiast if he had estimated that 15 per cent, or even 1 per cent, of granites were thus formed.

There is, therefore, some progress in our direction, and the recent discussions at the International Geological Congress in London have shown that we are not the only ones who think there are no magmatic granites in the strict meaning of the word. We admit, of course, that granites may result from the granitization of a lava, particularly of a rhyolite.

Professor Read states that he has yet to be convinced of the possibility of the abrupt stopping of metamorphism by diffusion on meeting porous layers. We do not see how he, as a "metamorphist," can otherwise explain the genesis of certain granites with strata of essentially unaltered quartzite within or in contact with the granites (Barrois, Lamotte and Routhier, Hietanen, *et al.*). The septa of limestone described by Moehlmann at the Ottawa meeting and many other phenomena also support this idea.

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VILLAFRANCHIAN STRATIGRAPHY IN SOUTHERN AND SOUTHWESTERN EUROPE[1]

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ABSTRACT

The deposits containing the classic Villafranchian fauna in Italy (Po Valley, upper and lower Arno Valley, and elsewhere in the peninsula), the Saône Valley of eastern France, and the Puy-en-Velay Basin (Haute-Loire) of central France not only consistently overlie sediments laid down during Upper Pliocene times but also are, in turn, immediately overlain by deposits that clearly belong to the first interglacial (Cromer Forest-Bed–Saint Prestian) stage. With the exception of peninsular Italy, where a marine regime still prevailed, the beds in question, of continental facies and belonging to a Villafranchian stage, represent a comparatively sudden break in the long sedimentary cycle that had prevailed in southern and southwestern Europe throughout Middle and Upper Pliocene (Plaisancian-Astian) times. During the Villafranchian stage, a marked lowering of the temperature occurred, supporting the view that the deposits laid down at this time were accumulated under conditions of glaciation in northern Europe, the Alps, and the Pyrenees. Furthermore, immediately prior to this phase of climatic change, a series of orogenic movements, apparently connected with a time of marine regression, profoundly modified the entire Apennine region. On the basis of these facts, the Villafranchian stage should not be retained in the Pliocene but should be considered basal Pleistocene.

INTRODUCTION

It is becoming increasingly apparent that no real progress can be made with regard to establishing a valid chronology as the basis for comparing the earliest finds of fossil man and of paleolithic implements in Europe and Asia, on the one hand, and the continent of Africa, on the other, until the status of the European deposits containing the well-known Villafranchian fauna is determined. It is generally admitted that in most of the Northern Hemisphere, at least, the beginning of the Pleistocene should be placed at the major climatic change which resulted in glaciation on a world-wide scale. Pilgrim (1944, p. 28), Zeuner (1945, p. 175; 1946, pp. 182–183), Flint (1947, p. 208; 1948, p. 546), and Cooke (1948, p. 43), the most recent authorities to discuss this problem, all seem to agree on this point. As regards the deposits themselves, however, there is as yet no widely accepted interpretation concerning what constitutes the actual stratigraphic base of the Pleistocene (i.e., what beds were laid down during the time of the earliest glacial stage) or, conversely, when the Pliocene came to an end.[2] In so far as Europe is concerned, these fundamental questions have not been adequately considered in the light of the sum total of the available field evidence since 1911, when Émile Haug published his *Traité de Géologie*.

At first glance it appears as though Haug's now-famous definition of the Pleistocene (Haug, 1911, p. 1767) were based entirely on paleontological evidence. However, careful analysis of his argument reveals that it was primarily on the basis of his extremely thoroughgoing and objective examination of all the available facts which led Haug to conclude that the climatic and geologic base of the Pleistocene coincided with certain paleontological data and that this hori-

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[1] Manuscript received January 28, 1949.

[2] For discussion and review of the status of the Pliocene-Pleistocene boundary problem in the light of Lyell's original definition and later work by Depéret, De Stefani, De Lapparent, Gignoux, and others see Flint, 1947, pp. 197–200; Schultz and Stout, 1948, pp. 571–574.

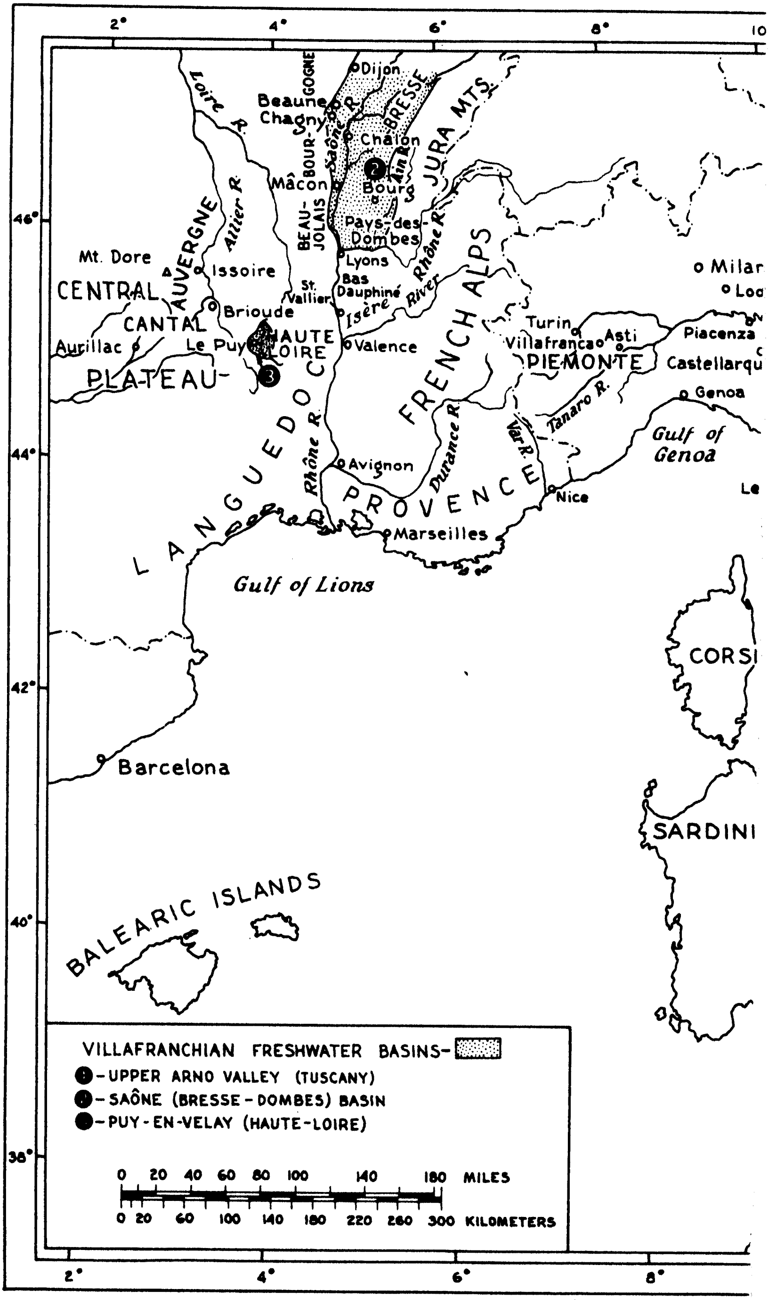
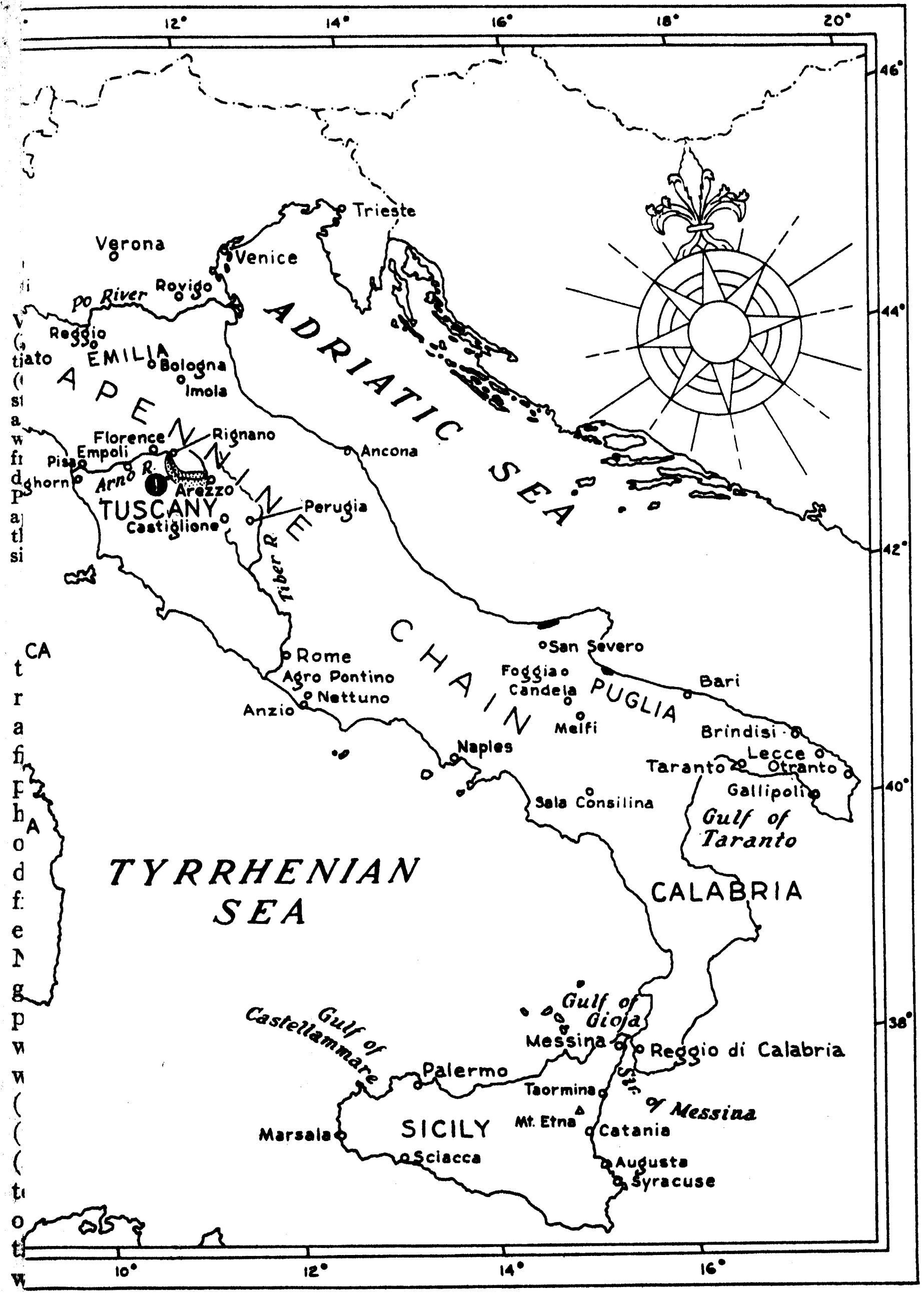


FIG. 2.—Map of southern and southwestern Europe, showing the location of Villa
Boule, 1892; Delafond and Depéret, 1893; Gignoux, 1913, pl. 5; and Merla, 1940.)



franchian and later localities mentioned in the text. (Based on Gaudin and Strozzi, 1859; Sacco, 1890;

zon was marked by the comparatively sudden introduction into Europe of the so-called "Villafranchian" fauna. Therefore, it is not the actual appearance of this fauna which Haug believed to be of fundamental importance; rather it is the fact that this fauna happens to occur at many localities in Europe in beds which this eminent authority was convinced were laid down during the first glacial stage of the Alps, Pyrenees, and Scandinavian regions. The evidence from Italy, Sicily, the Saône Basin of eastern France, and the Puy-en-Velay Basin of Haute-Loire to be considered in this paper[3] supports Haug's conclusion. It demonstrates not only that the Villafranchian represents a clearly defined stratigraphic unit overlying definitely Pliocene beds and is, in turn, overlain by deposits admittedly of the first interglacial stage, but also that, collectively, these deposits belong to a new cycle of sedimentation, probably reflecting the impact of climatic conditions colder and wetter than those now prevailing in southern and southwestern Europe.

THE PO AND ARNO VALLEYS AND OTHER LOCALITIES IN THE ITALIAN PENINSULA AND SICILY

The sequence of the various Pliocene and Pleistocene marine stages of the Mediterranean has been worked out in greater detail in Italy than anywhere else in southern Europe. Throughout this region the Pliocene is abundantly represented by thick marine deposits: clays accumulated in relatively deep water and sands laid down in coastal waters (fig. 1). Because the clay series, collectively known as the Plaisancian, is normally overlain by a sandy facies, grouped under the general term Astian, these formations have been regarded as representing two successive stages in the late Cenozoic succession of Italy. Investigations during the last few years, however, have clearly shown that this is not necessarily the case and that the marine clays and sands in question are merely different facies of the same stage (Gignoux, 1943, p. 528; Migliorini, unpub. MS). In the present paper, therefore, the two formations are referred to collectively as the "Plaisancian-Astian," although at many of the classic late Cenozoic localities in the Italian peninsula these deposits are normally superposed.

Of fundamental importance with regard to Villafranchian stratigraphy are the data which have been observed and studied at numerous exposures in the Po and Arno valleys and at several localities in southern Italy and Sicily. A summary of these observations and their stratigraphic significance follows.

PO VALLEY

Based on the detailed studies of Sacco, Pareto, De Stefani, Gignoux, and others in the southern portion of the Po Valley, the succession of the various late Cenozoic strata near Castellarquato (fig. 3) have become classic (cf. De Stefani, 1891, p. 326; Gignoux, 1915, pp. 338–345; 1916, pp. 26–35; 1923; 1943, pp. 550–552; Cushman, 1945). Here the deep-water blue clays of Plaisancian facies[4] are conformably overlain by the littoral or shallow-water yellow sands of the Astian;[5] the latter is, in turn, surmounted by the littoral sands with *Cyprina islandica* L., of the Calabrian. In 1865 Pareto[6] proposed Villafranca d'Asti as the type locality for the Villafranchian stage, and since then this term has been widely used with reference to those beds in southern and western Europe (as well as elsewhere in the Old World), which, on the basis of their stratigraphic position and contained fauna, have been assigned to this portion of late Cenozoic time. Most of the deposits of this stage consist of fluvial or fluvio-lacustrine sediments (fig. 2); at numerous localities these contain terrestrial and fresh-water fossils, including land snails. All along the northern Apennine slopes in the southwestern portion of the Po Valley the essentially continental Villafranchian sediments immediately overlie the marine clays and sands of the Plaisancian-Astian stage of the Pliocene epoch.[7]

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[3] Consideration of three important localities in the Allier Valley: Perrier, near Issoire; Mont Coupet, near Paulhaguet, and some 20 km. southeast of Brioude; and Senèze, located midway between Brioude and Paulhaguet, is omitted in this paper. Because the age of the deposits at Mont Coupet and Senèze can be established only on the basis of the interesting paleontological material found at these localities, the evidence of the deposits will obviously throw no light on the problem of Villafranchian stratigraphy. With regard to Perrier, lack of space prevents a discussion of the geological situation in the Issoire region, which is extremely complex.

[4] Plaisance, the type locality for this facies of the Pliocene in Italy, is a small village on the road

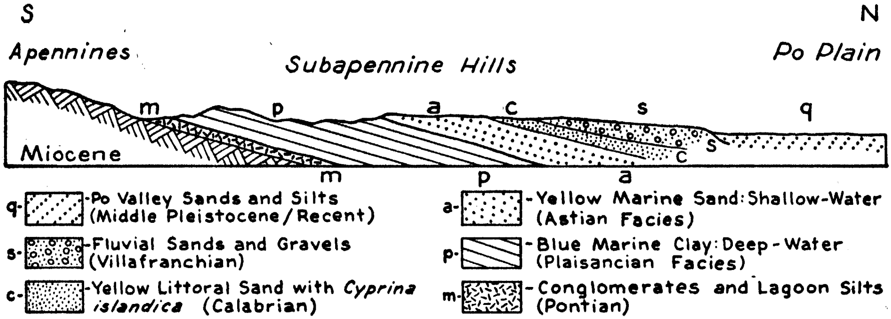
S
N
Apennines
Subapennine Hills
Po Plain
m
p
a
c
s
q
Miocene
s
c
m
p
a
q- Po Valley Sands and Silts (Middle Pleistocene/Recent)
s- Fluvial Sands and Gravels (Villafranchian)
c- Yellow Littoral Sand with *Cyprina islandica* (Calabrian)
a- Yellow Marine Sand: Shallow-Water (Astian Facies)
p- Blue Marine Clay: Deep-Water (Plaisancian Facies)
m- Conglomerates and Lagoon Silts (Pontian)

FIG. 3.—Schematic section showing the Pliocene and Pleistocene deposits in the vicinity of Plaisance, near Castellarquato, in the southern portion of the Po Valley. (After Gignoux, 1943, fig. 132.)

Although in the region of Asti the Villafranchian cycle of deposition is superposed directly on Pliocene sands, much of the rest of Italy, with the exception of the highland areas, was submerged. For the marine equivalent of the continental Villafranchian facies the term Calabrian, from the Province of Calabria in southern Italy, where deposits of this stage are extensive, was introduced in 1910 by Gignoux.[8] The invertebrate marine fauna of the Calabrian, studied in detail by Gignoux (1913, pp. 29–176), is interesting. In ad-

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between Castellarquato and Lugagnano. The term "Piacenzian," proposed by Gignoux in 1923, has never come into general use.

[5] Named after the interesting exposures in the region of Asti, Piedmont.

[6] Pareto, 1865, pp. 262–270. Villafranca d'Asti is a small town located on a tributary of the Tanaro River, some 15 km. due west-northwest of Asti; it lies on the railway line that runs between Asti and Turin.

[7] Depéret (1885*a*, p. 248; 1893, p. 528) points out the interesting fact that the Upper Pliocene deposits of southern France contrast markedly with those of northern Italy; for the evidence from both Montpellier and Roussillon (Perpignan), in the former region, demonstrates that a regime of fresh-water deposition was already well established by Upper Pliocene times. In Italy, however, this general movement was later, as proved by the sequence in the Asti region. After terrestrial conditions had become established in the Po Valley, the emergence of the Italian peninsula apparently progressed in a general north-to-south direction. Thus, in the south of Italy, marine sediments were being laid down contemporaneously with continental deposits in the north.

[8] 1910, p. 841; 1911, p. 115; 1913, p. 22. For the more or less transitional zone between the Calabrian

dition to an overwhelming majority of living species (over 95 per cent),[9] the Mediterranean was invaded during the time represented by these beds by a number of "cold" boreal types of mollusks from the North Atlantic. The new species, which suddenly appear here for the first time in Italy in Calabrian deposits, include the following:[10] *Cyprina islandica*, *Buccinum undatum*, *B. humphreysianum*, *Natica montacuti*, and *Trophon muricatus*. Although no studies have ever been made regarding the frequency of the occurrence of these forms with respect to the total assemblage of the marine fauna from any one of the main Calabrian localities, their appearance in the Mediterranean at this time is generally regarded as demonstrating markedly cooler marine conditions than those that had prevailed during Pliocene times.

This entire complex, comprising the continental Villafranchian and the marine Calabrian horizons, is included in the Pleistocene by Haug (1911, p. 1860), Rovereto (1925, p. 209), Blanc (1942*a*, p. 154), and Di Napoli-Alliata (1947, pp. 19 and 24). On the other hand, because Gignoux came across no exposures during the course of his extensive field work in Italy and Sicily that demonstrated an apparent sedimentary break between the Plaisancian-Astian and the Calabrian, he placed both these formations in the same sedimentary cycle. On this basis, Gignoux concluded that the Plio-Pleistocene boundary in Italy should be placed between the Calabrian and the next younger formation, the Sicilian. The results of recent investigations of this problem, however, have completely invalidated Gignoux's argument by demonstrating that the boundary in question coincides with the first appearance of northern immigrants in the faunistic succession of Italy, an event which occurred at the base of the Calabrian.

In the last few years several sections, composed in the main of marine deposits, have been recorded during the course of well-boring operations on the Po Plain (cf. Borghi, 1941; Di Napoli-Alliata, 1946*a*). According to Di Napoli-Alliata (1946*b*, p. 29), a series of deposits thus exposed near Rovigo clearly show the transgressive nature of the Calabrian. On the basis of the facts that (*a*) the basal Calabrian deposits here occur as a littoral formation, (*b*) the middle zone of the same horizon was laid down in relatively deep water, and, (*c*) finally, a series of littoral deposits demonstrate a new shallowing of the sea, Di Napoli-Alliata concludes that the Calabrian forms a complete sedimentary cycle. Furthermore, there is a most striking difference between the microfauna in the lowermost Calabrian deposits as contrasted with the Plaisancian-Astian beds (Di Napoli-Alliata, 1946*b*, p. 28). Approximately 10

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and the Villafranchian, characterized by brackish-water or lagoonal deposits, Sacco (1886) proposed the name "Fossanian"; but, because no characteristic fauna has ever been found for this zone, the term Fossanian is no longer used (see Gignoux, 1911, p. 120; 1913, p. 316). Apparently it represents the very base of the Villafranchian. (Cf. De Stefani, 1891, p. 322; Depéret, Mayet, and Roman, 1923, pp. 43-44.)

[9] As Gignoux points out (1922, p. 1453), the Calabrian deposits actually contain very few extinct species; those once thought to be extinct have in reality migrated to other regions, where they are still living.

[10] Gignoux, 1911, p. 117; 1913, p. 599; 1922, p. 1453. The whole question of the value of certain types of mollusks as indicators of late Cenozoic climatic conditions in the Mediterranean region is in urgent need of further investigation. In this connection, Dalloni (1940, p. 13, n. 3) reports the presence of *Cyprina islandica* L. in deep-water clays of Plaisancian facies on the Algerian coast, and Zeuner (1945, p. 232) records the same so-called "North Atlantic" form living at present in the Bay of Biscay as far south as the Bordeaux region.

per cent of the entire Calabrian assemblage is composed of cold forms, which thrive today in the North Atlantic, and these suddenly appear at the base of the formation. These are completely absent in the Plaisancian-Astian series. The significant quantitative differences revealed by this fauna have been summarized by Migliorini in a paper to be published in the forthcoming *Proceedings of the 18th International Geological Congress*, held in London in 1948.

From a second section near Lodi, southeast of Milan, Di Napoli-Alliata (1947, pp. 20–21) has described the plant remains found in a peat layer which is intercalated in a series of Calabrian clays and sands. Not only does this flora indicate that cool temperate conditions prevailed at the time of deposition, but also the Foraminifera from the same locality agree in every respect with this conclusion. As regards the flora, Di Napoli-Alliata states that, in place of the tropical to subtropical plants and trees that generally characterize the Pliocene deposits of this region, an essentially north-temperate assemblage was found, including *Alnus*, *Pinus*, *Abies*, *Castenea*, Ericaceae (a northern European family of heather), and various types of ferns. This flora indicates a climate somewhat cooler and wetter than that which prevails in northern Italy today. The subsurface deposits at the Lodi site also provide further evidence for a sudden shallowing of the sea at the base of the Calabrian as far west as the Milan region of the Po Plain.

South of Modena in the Apennine foothills fringing the southern edge of the Po Plain the transgressive nature of the Calabrian has also been observed at a newly excavated section near the town of Sassuolo. Here Montanaro-Gallitelli (1947, pp. 6–9) has described a clearly defined unconformity separating the Pliocene marine clays (Plaisancian facies), and the sands and gravels of the Calabrian series. This evidence confirms Ruggieri's conclusions based on his studies in the Santerno Valley, near Imola, some 70 km. southeast of Modena (Ruggieri, 1938, 1939, 1944). Thus, at a number of localities in the Po Valley the Calabrian beds have been recognized as being stratigraphically independent of the underlying Pliocene deposits, and there is a substantial amount of evidence demonstrating that the new cycle of sedimentation coincides with an abrupt break in the climate. But it is possible that the Calabrian itself covers more than one stage. Recent stratigraphic studies by Selli and Ruggieri in Emilia (Po Plain) favor the view that a new unit, Calabrian II or Emilian, with a mild-climate fauna, should be recognized between the Calabrian (*sensu stricto*) and the Sicilian (unpublished: paper presented at the 18th Internat. Geol. Cong., London, 1948; cf. Oakley, 1949*a*, *b*). However, the correlation between this new unit and the long interval of emergence accompanied by tectonic movements and erosion, which separates the Calabrian clays from the Sicilian yellow sands in the Imola region (Ruggieri, 1944, pp. 109–110), has not as yet been established.

Gignoux's detailed work on the Calabrian deposits of the southern portion of the Po Valley and along the east coast of Italy (Gignoux, 1915, 1916, 1922, p. 1485; 1923, pp. 374–375; also De Stefani, 1891, p. 322) has demonstrated that the replacement of this ancient northern Italian gulf by continental conditions progressed in a general west-to-east direction. Thus at Villafranca and numerous other localities in the Asti region, fresh-water deposition was already under way at the beginning of the Calabrian cycle of sedimentation. These latter deposits, which are mainly of fluvial origin, have yielded rich collections of the classic Villafranchian mammalian fauna.[11] Although the latter occurs throughout the widespread sands and gravels of this stage, which are preserved in the Asti region, the basal portion of these continental sediments is especially prolific of fossil bones. In fact, Depéret (1893, p. 531) states that at several localities proboscidian (*Archidiskodon planifrons* Falc.) molars with oyster shells attached to their surface have been collected (cf. De Stefani, 1891; also above, n. 8). However, with the exception of this single observation, one can learn little else concerning the nature of the actual gravel and sand beds which yield fossil vertebrate remains in this key area. Because much of this material was found in the last century when the excavation was being made for the bed of the railway between Asti and San Paolo di Villafranca, the stratigraphy should doubtless be rechecked by up-to-date methods. In any case, there appears to be no valid basis for the claim that bones and molars of true *Elephas* occur in this region in beds that are demonstrably of Pliocene age. The fact that in the vicinity of Imola abundant remains of a Villafranchian vertebrate fauna have been found in a sandy littoral deposit yielding mollusks typical of the Calabrian (Gignoux, 1915, p. 334; 1916, p. 22; 1923, p. 374; Ruggieri, 1938, 1939, 1944), on the other hand, provides further convincing proof that the Calabrian is definitely the marine counterpart of the essentially continental Villafranchian.

BASIN OF THE UPPER ARNO RIVER

During the last hundred and fifty years, enormous collections of fossil vertebrates have been amassed from various localities in the basin of the upper Arno River, where the late Cenozoic fresh-water sediments of this portion of Tuscany are well exposed. In point of fact, it is these deposits in the upper Val d'Arno which have furnished the type vertebrate fauna for the Villafranchian stage. This basin is filled with material that is entirely of fluvial and lacustrine origin (fig. 4) and which attains a thickness of over 350 meters. Its surface forms a broad, flat plain between Montevarchi, near Arezzo, on the south and Rignano-in-Valdarno in the north—a total distance of over 70 km., while to the south of Arezzo a long arm of this basin extends up the Val di Chiana to the vicinity of Chiusi, some 15 km. southwest of Castiglione (Merla, 1940; Losacco, 1944). There are two main series of deposits represented by the Val d'Arno Basin sediments, as follows:[12]

Lower series: Tilted beds of fine gray or blue lacustrine clays and silts (known to Italian geologists as *stellicione*), which contain fresh-water mollusks, layers of lignite, and a poorly represented mammalian fauna.[13] This horizon is of Plaisancian-Astian age.

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[11] For lists see Depéret, 1885*a*, pp. 256–257; 1893, pp. 530–531; Haug, 1911, p. 1861; Depéret, Mayet, and Roman, 1923, pp. 44–45.

[12] The writer is profoundly grateful to Professor Giovanni Merla, of Florence, for his generosity in providing an exceptionally clear statement of the results of his recent and as yet unpublished investigations in the Val d'Arno superiore. See also Gaudin and Strozzi, 1858–1859; Ristori, 1886*b;* Gignoux, 1913, pp. 304–309; Depéret, Mayet, and Roman, 1923, pp. 45–52; Sestini, 1936, pp. 38–39; Migliorini, unpublished paper to appear in the forthcoming *Proceedings of the 18th Internat. Geol. Cong.*

[13] The mammalian fauna from the *stellicione,* which includes *Anancus arvernensis, Sus provincialis, Tapirus* sp., and *Machairodus* sp., constitutes an assemblage without any well-determined character (cf. Depéret, 1893, p. 532). Apparently, this is the assemblage which Schlesinger (1922, p. 216) assigned to his "early" Val d'Arno horizon, his "late" horizon being the classic group mainly from the *sansino,* in which *Equus stenonis* Cocchi, *Leptobos etruscus* Falc., *Archidiskodon meridionalis* Nesti appear. Although the former is admittedly of Plaisancian-Astian age, Schlesinger's conclusions

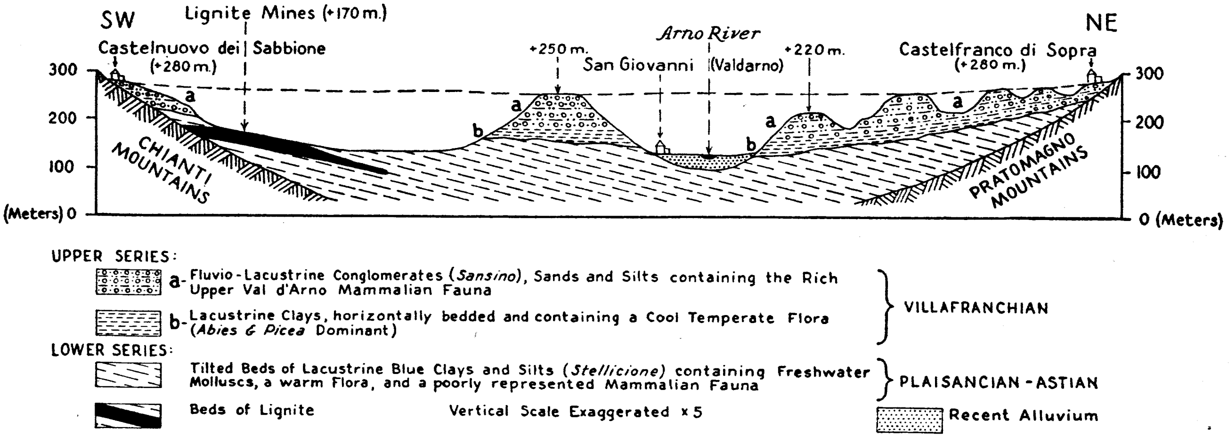
SW
Lignite Mines (+170 m.)
Castelnuovo dei Sabbione
(+280 m.)
+250 m.
Arno River
San Giovanni (Valdarno)
+220 m.
NE
Castelfranco di Sopra
(+280 m.)
300
200
100
(Meters) 0
0 (Meters)
a
b
CHIANTI MOUNTAINS
PRATOMAGNO MOUNTAINS
UPPER SERIES:
a- Fluvio-Lacustrine Conglomerates (Sansino), Sands and Silts containing the Rich Upper Val d'Arno Mammalian Fauna
b- Lacustrine Clays, horizontally bedded and containing a Cool Temperate Flora (Abies & Picea Dominant)
VILLAFRANCHIAN
LOWER SERIES:
Tilted Beds of Lacustrine Blue Clays and Silts (Stellicione) containing Freshwater Molluscs, a warm Flora, and a poorly represented Mammalian Fauna
PLAISANCIAN - ASTIAN
Beds of Lignite
Vertical Scale Exaggerated x 5
Recent Alluvium

FIG. 4.—Geologic section through the basin of the upper Val d'Arno, Tuscany. (After Gignoux, 1913, pl. 3, fig. 3; modifications based on Sestini, 1936, and Merla, unpublished ms.)

Upper series: These beds are horizontal throughout and are separated from those of the lower series by an unconformity. They may be subdivided into two clearly defined zones:

- *Zone b*—Lacustrine clays and silts, very similar to those of the lower series, comprise the basal member of the upper series. These clays are rich in pollen of trees which thrive under cool temperate conditions.
- *Zone a*—The fluvio-lacustrine gravels, or conglomerates (*sansino*[14]), micaceous sands and silts that occur throughout this zone, are of a ferruginous color. At many localities where these beds have been dissected and exposed by erosion, they are extremely rich in the remains of a typically Villafranchian vertebrate fauna.[15]

The character of the deposits of the upper series demonstrates that during Villafranchian times the late Cenozoic lake basin of the upper Val d'Arno was being transformed into a vast alluvial plain. According to Dr. Merla, all the polleniferous beds of Zone b occur below the level of the lowest strata in which the Villafranchian mammal remains have been collected. Because the beds of the upper series are practically horizontal throughout and because there is no evidence of an erosion interval during the time represented by the deposition of Zones b and a, it may be concluded that the mammal-bearing levels are everywhere more recent than those containing pollen. In any case this vertebrate fauna comes in the main from the uppermost 100 meters or less of the upper series—i.e., from Zone a and the top of the clays of Zone b.

It was on the basis of the evidence from the upper Val d'Arno that the Italian geologist, Sestini (1936), first demonstrated the fallacy of the concept of a stratigraphic continuity between the Villafranchian and the underlying definitely Pliocene beds. Sestini's work verifies the occurrence of pre-Villafranchian tectonic movements in this section of Tuscany and proves that the Villafranchian beds of the upper Arno Basin are separated from the older Plaisancian-Astian terrain by an unconformity. According to Sestini (1936, pp. 38–39), the strata comprising the upper (Zones a and b) horizon are nearly horizontal throughout, whereas in the lower (*stellicione*) series the bedding dips smoothly toward the north or northeast. In places the angle of tilt approaches 90°, but in the main it is not over 15°. A stratigraphic hiatus also occurs in the region immediately to the south; in the Val di Chiana, Losacco (1944, p. 65) has demonstrated that the Plaisancian-Astian beds are unconformably overlain by a thick series of lacustrine sands, clays, and gravels that contain a typically Villafranchian mammalian assemblage. The rich flora which has been documented from the layers of lignite in the upper part of the Plaisancian-Astian deposits of the upper Arno Valley is well known on the basis of the classic studies by Gaudin and Strozzi (1858–1859) and later by Ristori (1886), and it shows that tropical to subtropical conditions prevailed here during Pliocene times, just as elsewhere in southern and western Europe (cf. Sismonda, 1864,

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have been widely quoted as the authority for subdividing the Villafranchian stage as a whole into an "early" and a "late" horizon (cf. Zeuner, 1945, pp. 257–258; Schreuder, 1945, p. 190). Manifestly, no such subdivision is justifiable on the basis of the available data.

[14] *Sansino* is an Italian word that comes from *sansa*, the residue of crushed olives; that is to say, gravels containing elements up to the size of an olive (see Depéret, Mayet, and Roman, 1923, note on p. 46). The term is of no particular stratigraphic significance.

[15] For lists of the upper Val d'Arno fauna see: Depéret, 1885*a*, pp. 257–258; 1893, pp. 532–533; De Stefani, 1891, pp. 317–318; Forsyth Major, 1875, 1876, p. 345; 1885; Kormos, 1932; Depéret, Mayet, and Roman, 1923, pp. 50–53; Osborn, 1942, pp. 969–970; Schaub, 1928, 1943, pp. 286–287; Stehlin, 1934; Weithofer, 1889, 1891.

pp. 470–471; De Saporta, 1879, pp. 345–346; and Depape, 1928, pp, 47 and 67). Typical forms include *Sequoia*, *Taxodium*, *Liquidambar*, *Cinnamonum*, *Magnolia*, and *Cassia*, as well as other genera that thrive today in a warm climate. These forms contrast markedly with those represented in the spectra recently obtained by Dr. G. Merla (unpublished) from the peat beds intercalated in the Villafranchian lake sediments of Zone b of the upper series of this basin. The preliminary analysis of this pollen indicates the sudden spreading and rise to dominance of *Abies* (Fir) and *Picea* (Spruce) in this stratum, together with a rather high percentage of *Pinus* (typus *silvestris*) and substantial amounts of *Fagus*, *Alnus*, and *Tilia*. This assemblage provides convincing proof that the climate of the northern Apennines was perceptibly cooler than that of the present at the time these beds were laid down. Therefore, not only is the Plaisancian-Astian series of the upper Val d'Arno Basin stratigraphically and paleontologically quite distinct from the Villafranchian, but also the floras from these two deposits reveal that a marked climatic break occurred, which is suggestive of the onset of a glacial stage in the regions to the north.

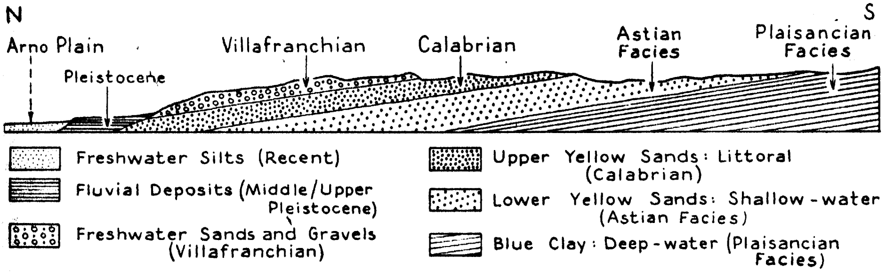
N
S
Arno Plain
Pleistocene
Villafranchian
Calabrian
Astian Facies
Plaisancian Facies
Freshwater Silts (Recent)
Fluvial Deposits (Middle/Upper Pleistocene)
Freshwater Sands and Gravels (Villafranchian)
Upper Yellow Sands: Littoral (Calabrian)
Lower Yellow Sands: Shallow-water (Astian Facies)
Blue Clay: Deep-water (Plaisancian Facies)

FIG. 5.—Diagrammatic section showing the sequence of late Cenozoic deposits in the lower Arno Valley, south of Pisa. (After Gignoux, 1913, pl. 3, fig. 2).

SITES IN THE LOWER ARNO VALLEY

As stated by De Stefani in 1891,[16] the late Cenozoic marine deposits of the lower Arno Valley, which belong to the Calabrian stage and which are overlain by fresh-water sediments, are closely parallel to and contemporaneous with the lacustrine deposits of the upper part of the same valley. This stratigraphic continuity between the fresh-water sediments of the upper Val d'Arno and the dominantly marine beds which extend from Empoli to the sea in the lower Val d'Arno agrees in all essential respects with the evidence summarized above from the Po Valley north of the Apennines (cf. Gignoux, 1916, p. 34); for, in that portion of Tuscany lying between Empoli and Pisa, the deep-water marine blue clays and silts of Plaisancian facies are overlain by a thick series of coarse yellow sands intercalated with gravel layers (fig. 5). Although no recent work has been done in this region, the basal part of the latter horizon is considered to belong to the Astian series, whereas the upper levels have been assigned to the

[16] De Stefani, 1891, pp. 314–317; see also Forsyth Major, 1885, pp. 3–4; Gignoux, 1913, pp. 300–304 and 307–308; 1922, pp. 1464–1465; Depéret, Mayet, and Roman, 1923, pp. 46, 52–53, and 116–118.

Calabrian.[17] Especially in the eastern portion of this region, a considerable thickness of fresh-water silts, sands, and gravels overlies the marine sediments of the Calabrian. The latter mark the final regression of the sea from the valley and in part represent a synchronous prolongation of the uppermost fluvial and fluvio-lacustrine deposits of the upper Val d'Arno.

As is universally the case throughout the latter area, important finds of a typically Villafranchian vertebrate fauna[18] have been made both in the fresh-water gravels and in the upper part of the immediately underlying marine sands. In all its main elements this fauna is identical with that from the upper Val d'Arno, a fact which, together with the geological evidence, provides absolutely no basis for the argument that the mammalian assemblage of the upper Val d'Arno is more recent than that from the lower Val d'Arno. Certainly in no respect whatsoever do the data support Boule's claim for assigning a pre-Villafranchian age to the deposits in the lower Arno Valley containing the *Archidiskodon meridionalis* fauna (Boule, 1893*c*). As far as Italy is concerned, the stratigraphic relationships between the Villafranchian-Calabrian complex and the next younger marine stage, known as the Sicilian,[19] is exceptionally clear. This is especially true in the case of many fine sections in the Leghorn and Anzio regions, as well as in southern Italy and Sicily, where very thick deposits of this stage are exposed.

SOUTHERN ITALY AND SICILY

As demonstrated by the evidence from the upper Val d'Arno, the transgressive nature of the Calabrian on the Po Plain is in accord with the fact that mountain-building processes took place between Plaisancian-Astian and Calabrian times in the northern Apennines. That severe orogenic movements did, in fact, occur at this time throughout the entire Apennine chain is proved by Migliorini's recent observations in the Candela-Melfi region, south of Foggia. Here, in the Apennine foothills of Puglia, the Plaisancian-Astian series is bent into recumbent folds, whereas just to the east there occur Villafranchian beds which are practically horizontal. Furthermore, near Sala Consilina, some 76 miles due southeast of Naples, the Apennine divide is capped by a marine Plaisancian-Astian outlier at an elevation of some 1,150 meters above sea-level (Migliorini, 1946). At the base of the scarps on either side of this watershed there is a series of basins, each of which is filled with Villafranchian lacustrine deposits, their surfaces ranging from 650 to 800 meters above sea-level. In spite of this very clear stratigraphic evidence, however, no true unconformity is perceptible between the Plaisancian-Astian and the Calabrian either in the classic region of Calabria (Migliorini, unpub. MS) or in the Messina area of Sicily (Trevisan and Di Napoli-Alliata, 1937, p. 33; Trevisan, 1942, p. 11). It was on this evidence that Gignoux assumed conformity at the many other localities in Italy and Sicily which he studied between 1908 and 1913. Nevertheless, as originally pointed out in the vicinity of Sciacca by Di Stefano (1889, pp. 71–72) and later near Augusta by Maugeri-Patané (1923, pp. 94–95), in the Syracuse region by Trevisan (1936, pp. 10–11), and at several additional localities in southern and southwestern Sicily (Cipolla, 1934*a;* 1934*b*, p. 3; Trevisan and Di Napoli-Alliata, 1937, p. 34; Blanc, 1942*a*, p. 149; Trevisan, 1942, pp. 8–9), there is a very well-defined angular unconformity at the base of the Calabrian which clearly separates it from the underlying Plaisancian-Astian complex. Also, the presence of this well-marked stratigraphic feature at Monte Mario, near Rome (Cerulli-Irelli, 1907, pp. 67–68; Haug, 1911, p. 1645; Blanc, 1942*a*, p. 149), has been established.

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[17] Demonstrated by Gignoux (1913, pp. 303–304) on the basis of his detailed study of the invertebrate marine fauna from several late Cenozoic horizons exposed at Vallebaija and other localities in the vicinity of Leghorn. In this connection, it should be noted that remains of *Archdiskodon meridionalis* have been collected in the Calabrian deposits exposed in the vicinity of Fauglia, near Vallebaija (see Gignoux, 1913, pp. 308 and 614).

[18] For lists see: Depéret, 1893, p. 531; Depéret, Mayet, and Roman, 1923, pp. 53–54 and 118.

[19] For a clear definition of the Sicilian, a stage originally described by Doderlein (1870–1872), see Gignoux, 1908; 1911, pp. 121–123. According to the stratigraphic scheme advocated by Gignoux, the Sicilian marks the base of the Pleistocene. Osborn (1910, pp. 317–321; 1915, p. 222), as well as Boule and Piveteau (1935, pp. 516 and 561), following Doderlein (1870–1872), Seguenza (1880, pp. 315–317), and De Lapparent (1906, p. 1648), group both the Villafranchian-Calabrian *and* Sicilian stages under the general term "Sicilian," which they assign to the Pliocene (cf. Gignoux, 1913, pp. 7–9; Cipolla, 1934*b*).

At numerous localities in Sicily the transgressive nature of the Calabrian is especially clear, confirming the evidence from the Po Plain, summarized above. From Marsala in southwestern Sicily, all along the south coast, and as far north as Augusta in the eastern portion of the island the Calabrian rests on Pliocene and older rocks, and there is an angular unconformity between this complex and the deposits of the Plaisancian-Astian series. Trevisan (1942, pp. 3 and 8–9) states that during the time immediately preceding the lower Calabrian there was a major phase of uplift followed by an important marine transgression. Although this event is clearly recorded in the Syracuse area, the Messina region forms a special belt in which the discordance between the Pliocene and the Calabrian horizons has not been registered. Because the transgressive portion of the Calabrian series occurs at approximately the level of the present shore, it is concluded that the pre-Calabrian regression caused a somewhat greater extent of the land to be emerged than is the case at present.

The strata exposed at several localities along the western coast of the Italian peninsula—Leghorn region (Gignoux, 1911, p. 126; 1913, pp. 309–313), the coastal plain of Agro Pontino, south of Rome, and between Anzio and Nettuno (Blanc, 1937*a*, p. 629)—clearly reveal the stratigraphic position of the 80–100-meter raised beach deposits of the Sicilian stage in relation to the underlying sediments of the Calabrian-Villafranchian complex (cf. Blanc, 1937*b*, p. 359; 1942*a*, p. 154). In the Anzio area (Blanc, 1937*a*, p. 629) the Calabrian sandstones were "dislocated during the Lower Pleistocene by tectonic movements and partly destroyed by degradation. . . . The post-Calabrian regression was followed by the Sicilian transgression, during which the sandy clays with *Plicatula* and *Brocchia* to the south of Nettuno were laid down in roughly horizontal strata." It is this evidence which serves to connect the late Cenozoic sequence in northern Italy with that recorded at many of the classic localities of the southern portion of the peninsula and Sicily (fig. 6). In each case the Sicilian deposits directly overlie those of the Calabrian, although the two are stratigraphically separated from each other by an unconformity represented by tilting and erosion. In other words, during the interval between the Calabrian and the Sicilian a regression of the sea, accompanied by a new phase of tectonic movements, occurred.

In the Rome region, gravels of Sicilian age are widespread, as pointed out by

Blanc (1948, p. 5). Two localities near the city of Rome have yielded very early lower Paleolithic implements, as follows: (*a*) a heavily rolled hand-axe of Abbevillian type (G. A. Blanc, 1935) and (*b*) two flakes manufactured by the so-called "Clacton" technique (A. C. Blanc, 1936). Although the horizon in question is apparently the same in each case and is believed to date from Sicilian times (cf. Köppel, 1935, p. 475; Blanc, 1942*b*, p. 5; 1948, p. 9), this fact has not as yet been definitely established.

According to Migliorini (unpub. MS), the orogenic movements between the Calabrian and the Sicilian were very much less severe than those which occurred during the Plaisancian-Astian-Calabrian-Villafranchian interval. In the Apennines very mild mountain-building activity is evidenced by occasional minor tilting and faulting of the Villafranchian lacustrine deposits. Trevisan (1942, pp. 16–17) states that along the coast a vast "Tirrenide" appeared at this time, which was almost immediately transgressed by the sea. But very probably eustatic factors were also involved. In addition to the evidence cited above from the Anzio-Nettuno area, there is a clearly defined unconformity between the Calabrian and the Sicilian in the Palermo and Agrigento regions (Trevisan, 1942, pp. 20–21; Cipolla, 1934, p. 6), as well as elsewhere in Sicily, e.g., Mortelle, near Messina (fig. 6).

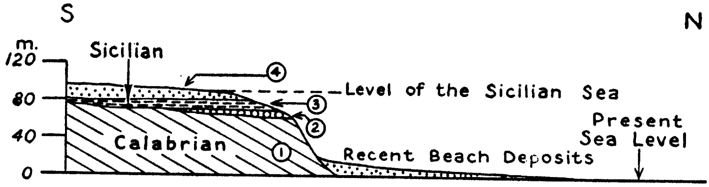

FIG. 6.—Section through the marine deposits in the vicinity of Mortelle, near Messina (Sicily), showing the unconformity between the Sicilian raised beach and the underlying Calabrian strata. *4*, littoral marine sands, very fossiliferous; *3*, beds with *Ostrea edulis; 2*, conglomerates and lagoon silts containing *Cardium edule; 1*, very hard conglomerates with large *Mytilus* shells. (After Gignoux, 1913, fig. 31.)

Gignoux's detailed observations in Calabria and Sicily (1908, 1909, 1910, 1911, 1913; summarized by Depéret, 1918, p. 483) demonstrate that the invertebrate marine fauna of the Sicilian stage, the type locality for which is in the region of the Gulf of Palermo, is intimately and directly related to that of the Calabrian deposits of the same area.[20] Recent studies of the mollusks from Sicilian clays at Ficarazzi and Conca di Oro, near Palermo, have further substantiated this conclusion (Tamajo, 1937, p. 466; De Stefani, 1941, p. 280), while the Foraminifera from a new site in southwestern Sicily, where the deposits are carried on an extensive wave-cut terrace that is up to 3 miles wide, actually differ very little from those recorded at Calabrian deposits in the same region (Trevisan and Di Napoli-Alliata, 1937, p. 25). Normally, however, the Sicilian assemblages are characterized by (*a*) the disappearance of a few of the "archaic" Pliocene types which persisted in the Calabrian; (*b*) the presence of some, very rare, extinct forms; and (*c*) the occurrence of a limited number of "cold," or boreal, species typical of northern Atlantic waters, most of which had already appeared in Calabrian times. This fauna is found in two types of sediments—true beach accumulations (fig. 6) and ocean-floor deposits—both of which were formed at a time when the sea relative to the land stood some 80–100 meters above its present level (cf. Depéret, 1920, p. 163). As previously stated, these Sicilian marine gravels, sands, and silts rest unconformably on the older Calabrian series. But the erosion interval separating the two horizons must have been of relatively short duration (Trevisan, 1942, p. 22), since, from a paleontological point of view, the dominantly littoral faunas of the Sicilian and Calabrian stages are very closely related. Possibly, further light will be thrown on this problem if deep-sea cores covering this portion of late Cenozoic time can be obtained for analysis. In any case, there is no evidence whatsoever of an uplift during Sicilian times.

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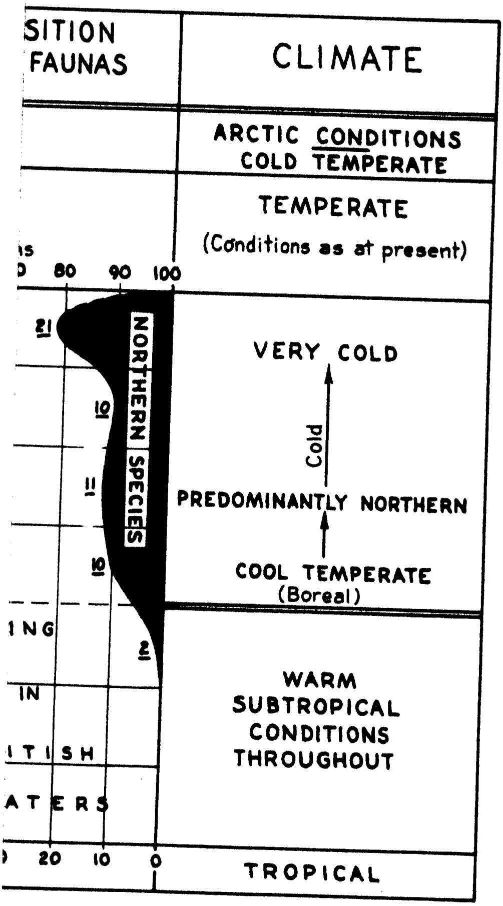
[20] During Plaisancian-Astian times, a large part of the island of Sicily was under the sea (Gignoux, 1922, p. 1474). Toward the close of the Pliocene, however, deposition was taking place under conditions of gradual uplift, a movement which culminated in the pre-Calabrian emergence (Trevisan, 1942, p. 8). As the result of the subsequent transgression, marine deposits of the Calabrian stage are enormously thick even in central Sicily (Gignoux, 1911, pp. 119 and 125; Trevisan, 1942, pp. 9–16), which is also the case in southern Italy.

The evidence concerning the climate of the Sicilian is confusing. Formerly, on the basis of the "cold" increment in the molluscan faunas of the Sicilian, it was believed that the temperature of the sea during this stage was actually somewhat lower than that of the present Mediterranean. This conclusion is supported by studies of the microfauna from the classic deposits at Conca d'Oro, near Palermo (Di Napoli-Alliata, 1937). On the other hand, most of the so-called "boreal" types, once thought to be new arrivals from the North Atlantic during the Sicilian, apparently were already present as early as Calabrian times. As Gignoux (1922, p. 1475) points out, the forms considered to be indicative of warm conditions are very much more abundant in deposits of this stage than are the typically "cold" assemblages. Furthermore, many localities yield a rich fauna that is neither characteristically "cold" nor characteristically "warm," because the total number of species that serve as critical climatic indicators is very small in comparison with the fauna as a whole. It is hoped, therefore, that some day the data will be available to provide a basis for working out the percentage composition of the molluscan faunas from some of the main southern Italian and Sicilian localities of this stage, just as Davies has done in the case of the Pliocene and Pleistocene deposits of southeastern Britain (cf. fig. 7; Davies, 1934, fig. 28, p. 208; also Boswell, 1936; Zeuner, 1937, 1945, p. 106). With reference to those sections in southern Italy and Sicily at which the so-called "cold" Sicilian fauna has been recorded, Depéret (1918, p. 483) states that doubtless the deposits in question were laid down in certain cold currents at a given depth in the Sicilian Sea. According to this authority, the "cold" forms are completely lacking at many other Sicilian localities in Italy, along the southern coast of France, and in North Africa, where the invertebrate fauna corresponds in all its essential elements with a normal Mediterranean assemblage of the present day.

On the basis of the recent data, Migliorini (unpub. MS) considers it very doubtful that any cooling at all took place in the Sicilian. Indeed, the very meager paleobotanical evidence from near Lodi, southeast of Milan on the Po Plain (see p. 384; Di Napoli-Alliata, 1947, p. 22), strongly suggests that a significant *rise* in temperature may have occurred during early Sicilian times. This



e faunas. (Based on Davies, 1934, fig. 28; also Osborn,
1.)

N C E PUY-EN-VELAY BASIN (HAUTE-LOIRE)	VERTEBRATE PALAEONTOLOGY	CLIMATE
IPPER SANDS AND GRAVELS SAINZELLES & LA MALOUTEYRE	CROMER FOREST-BED/ SAINT-PREST FAUNA with *Elephas* (*Hesperoloxodon*) *antiquus*	Conditions very Similar to those Prevailing at Present
Erosion Interval LUVIAL SANDS, GRAVELS & CLAYS TERCALATED WITH BASALTIC FLOWS ASTODON SANDS OF PUY elay Fauna: Vialette, Crozas, etc. eyssac: Diatomaceous Clays with Plant Remains	VILLAFRANCHIAN FAUNA ELEPHAS — *Archidiskodon* Group EQUUS — Zebrine Group LEPTOBOS — Primitive Cattle	CLIMATE COLDER AND WETTER THAN AT PRESENT First Stage of Continental Glaciation
Prolonged Erosion Interval VOLCANIC ERUPTIONS CONTINUED asalts, Andesites, Phonolites etc. sh Layers with Plant Remains	MONTPELLIER AND ROUSSILLON (PERPIGNAN) FAUNAS Mammalian Faunas of this Stage are very poorly Represented in Europe	WARM CLIMATE Tropical to Subtropical Conditions
FIRST PHASE OF VOLCANIC ACTIVITY ASSOCIATED WITH TECTONIC MOVEMENTS	HIPPARION-FAUNA	SAVANNAH CLIMATE Tropical to Subtropical Conditions

western Europe

is also indicated by the complete absence of northern species of Foraminifera collected by Ruggieri (1944, pp. 109–110) in the Imola region, southeast of Bologna. The discovery of a rich polliniferous deposit in a definitely Sicilian context would provide the necessary data to solve this problem. In the meantime, the evidence to date certainly does not substantiate the argument that climatic conditions in Italy during Sicilian times were markedly cooler than those of the present day.

At three localities in Sicily—Conca d'Oro di Palermo, in the vicinity of Taormina at the base of Mount Etna, and Mortelle (fig. 6), a small village near Messina (Gignoux, 1911, p. 128; 1913, pp. 615–618; 1922, p. 1478; Depéret, Mayet, and Roman, 1923, pp. 171–172)—this horizon, which is represented in the Messina region by a littoral conglomerate cut into the inclined Calabrian beds, has yielded remains of *Elephas* (*Hesperoloxodon*) *antiquus* Falc. Other finds of this ancient, African type of elephant have likewise been recorded from Sicilian deposits on the Italian mainland—near Reggio di Calabria (three localities), in the Leghorn region, and at San Severo, near Foggia.[21] Since the San Severo find, a very typical M_2 of *E. antiquus*, was discovered below a thickness of some 14 meters of marine gravels of the Sicilian stage, it is difficult to avoid the conclusion that the latter is of Lower Pleistocene (first interglacial) age. In any case, there seems to be absolutely no basis for the contention that the Sicilian stage should be placed in the Pliocene epoch.

THE SAÔNE BASIN OF EASTERN FRANCE

In the Saône Basin of eastern France, which is comprised of the Bresse region and the Pays-des-Dombes, the late Cenozoic stratigraphic sequence was worked out in great detail by Delafond and Depéret. Although several important contributions have subsequently appeared, the main succession of five Pliocene and Lower Pleistocene geological and paleontological events, as established by these eminent authorities, still remains substantially as they set it forth in their classic monograph, published in 1893 (cf. Béroud, 1909; Roman, 1929, p. 138; Pelletier, 1946*a*, pp. 131–138). As in the upper Arno Valley, the late Cenozoic deposits of this area consist of two main series of deposits, and these occupy nearly the entire extent of the basin. The older series is almost exclusively lacustrine, whereas the younger beds were for the most part laid down under fluvial conditions (fig. 2).

Between the Lower Pliocene freshwater sands, which yield a very typical Pontian (*Hipparion*) fauna (Depéret, 1887, pp. 104–112; 1894*c*, pp. 715–718; Delafond and Depéret, 1893, pp. 41–56; Viret, 1947), and the overlying Middle Pliocene lacustrine sediments, there is a marked disconformity caused by uplift and erosion. But the Middle Pliocene was everywhere a stage of depression and marine dominance, when the sea extended up the Rhône Valley to the vicinity of Givors within 20 km. of Lyons (see Fontannes, 1879–1882, especially the large folding map in vol. 2) and the entire Bresse-Dombes region was occupied by a lake,[22] as shown in figure 1. The beds laid down in the latter area at this time contain a rich molluscan fauna of Levantine facies, as well as a mammalian assemblage[23] composed in the main of new invading types of animals not found in the earlier deposit.

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[21] Gignoux, 1913, pp. 280 and 615; Depéret, Mayet, and Roman, 1923, p. 172. The abundant remains of *E. antiquus* Falc. from the Sicilian deposits of the Leghorn region are stated to occur in association with *Hippopotamus*.

[22] This Middle Pliocene Lake of Bresse, which was over 200 km. long, emptied into the narrow marine "ria" of the Rhône Valley south of Lyons (see Tardy, 1883, pp. 562–564; 1886, pp. 91–101; Depéret, 1895, p. 445; Denizot, 1934, p. 642;

At the end of the Middle Pliocene a new phase of orogenic movements connected with mountain-building processes led to the partial withdrawal of the sea from the Rhodanian Gulf and an erosion interval in the Bresse-Dombes Basin. In the valleys cut by the rivers at the expense of the Middle Pliocene lacustine beds during this interval, a dominantly fluviatile facies—sands, gravels, and tuffs—was accumulated (Delafond and Depéret, 1893, pp. 180–185). At Trévoux, an important locality of this stage on the Saône River, typically Upper Pliocene mammalian remains, mollusks, and imprints of leaves (Delafond, 1885; Depéret, 1889, 1894*b*, pp. 625–626) have been recorded, and similar deposits occur elsewhere in the region. The fossil vertebrates from these beds[24] have direct affinities with the well-documented contemporary assemblages of Montpellier and Roussillon (Perpignan) in southern France, the classic localities for Dépéret's "Astian fauna." In the lower valley of the Ain, in the extreme southeastern portion of the Dombes region, a tremendously rich flora of this stage (Delafond, 1886*a;* Delafond and Depéret, 1893, pp. 186–187; Depéret, 1894*a;* Roman, 1936*a*, p. 162; 1936*b*, p. 186) has been described in detail from the locality of Meximieux.[25] These Meximieux plant beds (fig. 9), which have become classic, were laid down under conditions of a warm, subtropical climate, when the prevailing conditions in the Bresse-Dombes Basin were very similar to those obtaining at present in the lower Mississippi Valley. With regard to southeastern France, Depape (1928, p. 81) has calculated that, during Upper Pliocene times, the 20° C (68° F.) isotherm passed through the general region of Avignon, while in the slightly more northerly Bresse-Dombes Basin the annual mean was between 17° and 18° C. (62°–64° F.). Thus the thick Upper Pliocene sand deposits, pebbly beds, and tuffaceous layers of the Saône region were accumulated during a period when the climate in the basin averaged some 5°.5 or 6° C. warmer than it does at present.

The next series of beds, which belong to the Villafranchian stage, appear to represent an abrupt break in the sedimentation of the Bresse-Dombes region. The characteristic deposits consist of very much coarser sediments; these occur in the form of huge terrace and fan formations, which are found in all the main valleys. They issue from the flanking massifs—those of Beaujolais and Bourgogne on the west and the Jura on the east—and disconformably overlie the older terrain.[26] These fluvial sands and

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Roman, 1936*a*, p. 160). As Gignoux (1943, p. 555) has observed, the situation during Middle Pliocene times in the Saône Basin is very reminiscent of the conditions which prevailed in the intra-Apennine depressions of Italy; the latter were also occupied by lakes during an otherwise dominantly marine stage.

[23] For lists see Depéret, 1893, p. 534; Delafond and Depéret, 1893, pp. 124–125; Viret, 1939; Pelletier, 1946*b*.

[24] For lists see Depéret, 1893, p. 535; Delafond and Depéret, 1893, pp. 191–197.

[25] De Saporta and Marion, 1872; De Saporta, 1873, pp. 217–219; 1879, pp. 322–337; Depape, 1928, p. 47. According to Laurent (1911, p. 90), an identical flora, also of Upper Pliocene age, occurs in a series of tuff deposits exposed not far from Belley (Ain), in the valley of the Furans River, a tributary of the Rhône.

[26] Delafond, 1886*b*, p. 69. For these widespread and very characteristic deposits of the Saône Basin the term "Bressian Conglomerate" (*Conglomérat*

gravels, which Depéret and Penck considered to be the equivalent of the *älterer Deckenschotter* of the northern Alps (Depéret, 1898, p. 423; Penck and Brückner, 1909, p. 648), are very extensive; they not only cover almost the entire Bresse-Dombes Basin (fig. 10) but are also found throughout the valley of the Rhône River and its tributaries south of Lyons.[27]

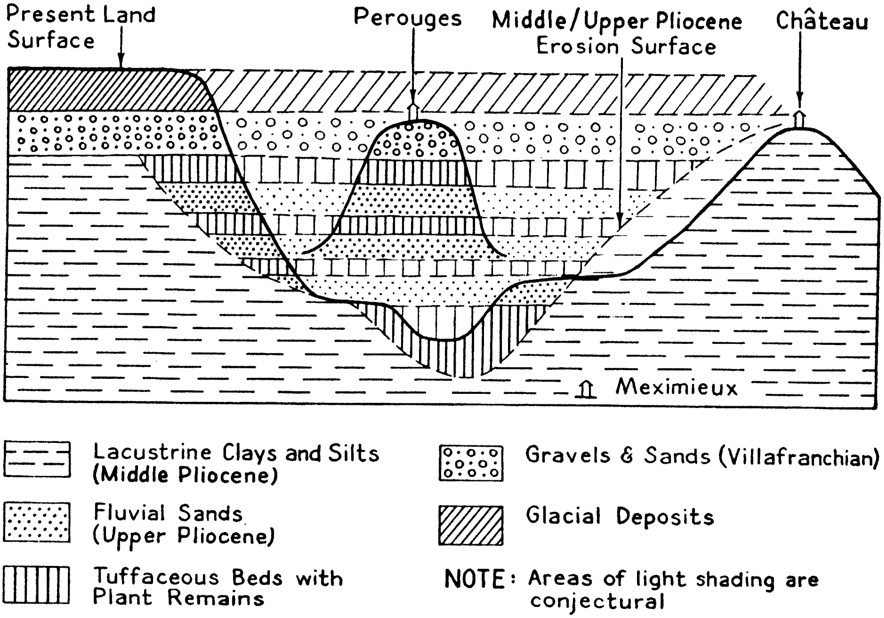
Present Land Surface
Perouges
Middle/Upper Pliocene Erosion Surface
Château
Meximieux
Lacustrine Clays and Silts (Middle Pliocene)
Fluvial Sands (Upper Pliocene)
Tuffaceous Beds with Plant Remains
Gravels & Sands (Villafranchian)
Glacial Deposits
NOTE: Areas of light shading are conjectural

FIG. 9.—Schematic section of the deposits near Meximieux (Ain), showing the stratigraphic position of the Upper Pliocene tuffaceous beds containing abundant plant remains. (After Delafond and Depéret, 1893, fig. 43.)

Along the foothills of the southern Jura, from Bourg-en-Bresse south to the Rhône Valley, these fluvial deposits, which contain numerous pebbles of Alpine origin up to 20 cm. in diameter, form a more or less continuous sheet (Depéret, 1885*c*, p. 122; 1898, p. 422; Delafond, 1889, pp. 7–8; Boistel, 1898*b*, pp. 57–59). Near Ambérieu, a town located 12½ km. northeast of Meximieux, several fine exposures of the gravels of this stage have been studied in detail by Boistel (1894, p. 300; 1898*a*; pp. 31–33; 1902, pp. 138–139), who states that, although they have been much reduced by erosion, they once attained a thickness of over 40 meters at several places. Indeed, Depéret (1885*c*, p. 123) records a section near Montluel (see fig. 1), where he studied a vertical

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Bressan) was originally proposed (cf. Benoît, 1858, p. 328; Fontannes, 1885, p. 60; Depéret, 1888), but it was subsequently abandoned.

[27] The Plateau Gravels, or *Cailloutis des Plateaux*, of the Lyons region and Bas Dauphiné (forests of Chambaran and Bonnevaux), as well as elsewhere in the lower Rhône Valley, which are also of Villafranchian age, likewise constitute an immense surface of alluvial deposits that has been described by many authors (cf. Fontannes, 1884; Riche, 1886, 1887, 1888; Depéret, 1895, p. 445; 1913, pp. 533–534; Kilian, 1902, pp. 160–161; 1911, p. 98; Martin, 1905–1906, pp. 146–157; De Lamothe, 1906, p. 1104; 1910 pp. 806–807; 1915, pp. 70–87; 1921, pp. 105–

section of these gravels approximately 80 meters high. According to this latter authority, it is evident on the basis of the thickness and extent of this vast sheet that it represents the eroded remnant of an alluvial terrace of the ancient Rhône which is also Viret's (1926, p. 55; 1931,

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114; Kilian and Gignoux, 1910, pp. 1023–1024; 1911, pp. 203–210; Briquet, 1911, pp. 37–39; Faucher, 1913; Kilian and Révil, 1917–1918, pp. 213–235; Denizot, 1933, pp. 560–573; 1934, pp. 615–620; 1939, p. 125). These thick *Elephas* (*Archidiskodon*) *meridionalis*-bearing gravels, regarded by Penck (with Brückner, 1909, pp. 649–659; see also Schaudel, 1907, pp. 139–155) as corresponding in the main with the *älterer Deckenschotter* of the region to the north of the Alps, presumably are remnants of the outwash fans of a glaciation, the moraines of which have been long since dismantled as the result of subsequent erosion and extensive glaciation. Indeed, as Bourdier has recently stated (1946, p. 339; see also Depéret, 1922, p. 1419), the existence of actual morainic deposits attributable to the Günz stage in the French Alps is a very problematic matter. At the locality of La Croix-Thoré in the commune of Saint-Vallier (Drôme), however, the Plateau Gravels appear to be associated with an ancient calcareous loess (Bourdier, 1942, p. 423; 1947) from which Viret has recovered a very rich assemblage of mammalian remains, including *Equus stenonis*, *Crocuta* cfr. *perrieri*, and numerous species of typically Villafranchian artiodactyls. Not only does this evidence suggest that a Villafranchian fauna was living in southeastern France under the periglacial conditions of the Günz glacial stage, but also that the Plateau Gravels (*Cailloutis des Plateaux*) should be included in the Pleistocene.

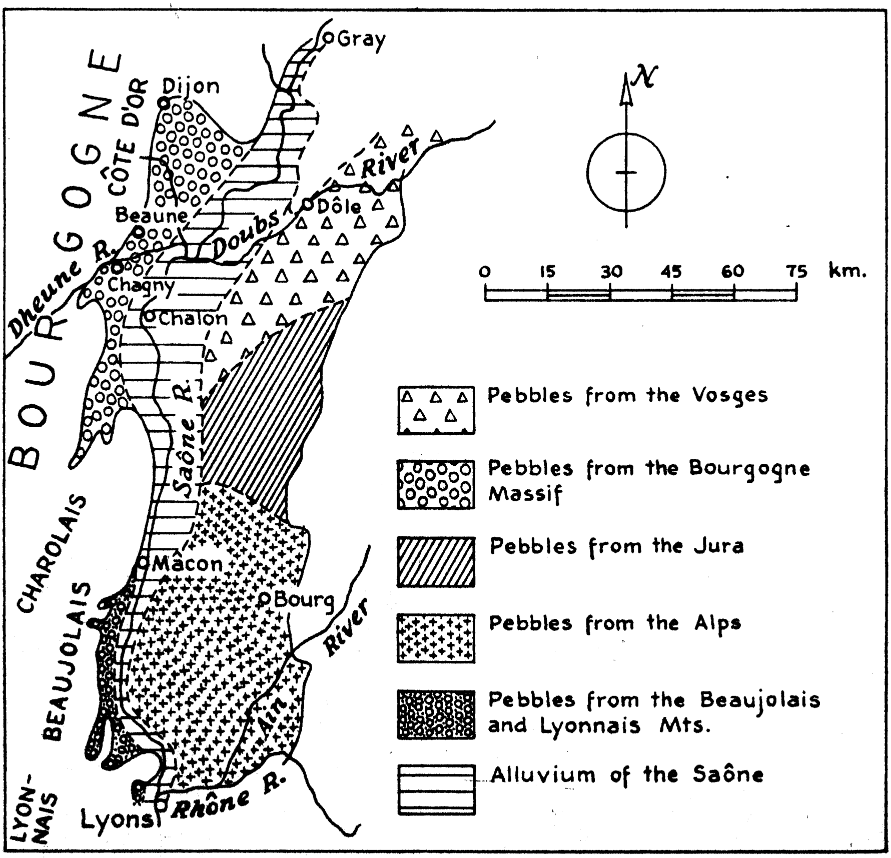

FIG. 10.—Map showing the distribution of the Villafranchian gravels and the source of their main constituents in the Saône (Bresse-Dombes) Basin. (After Delafond and Depéret, 1893, fig. 43.)

p. 168) opinion. This was therefore an intensive and widespread stage of alluviation, when all the streams issuing from the highlands were rapidly eroding the beds of their upper courses and transporting this material to the adjacent lowlands and plains, where it was deposited. As Delafond and Depéret (1893, p. 202) point out, this was an unprecedented period of erosion and deposition, which far exceeds in importance similar events associated with subsequent, and demonstrably Pleistocene, stages in the highlands of the western Alps and those of the Jura.[28]

It seems probable that this abrupt change in the depositional processes operating in this region is connected with a major climatic break that resulted in the first advance of the Alpine ice-sheets. Journaux (1948, p. 70) has recently reported definite signs of cryoturbation (*or* congeliturbation, see Bryan, 1946, p. 633; 1948) caused by frost action contemporaneous with the deposition of the huge outwash fans of this stage found along the base of the eastern flanks of the Côte d'Or, south of Dijon. Furthermore, in these fresh-water Villafranchian sediments, the constituents of which consist predominantly of erratics from the Vosges, Jura, and Alpine areas (fig. 10), angular and relatively unrolled boulders of Alpine origin, up to 50 cm. in diameter, have been reported (Boistel, 1902, p. 157; Journaux, 1948, p. 69). This evidence demonstrates that the deposits in question were accumulated under torrential conditions, when the climate was colder than it is at present in this region and when the transporting power of the streams was greatly in excess of what it has been during Recent times.

Now the relative age of these high-level gravel and sand deposits is established beyond all cavil on the basis of the very clear paleontological evidence from a great number of localities, where remains of characteristic Villafranchian mammals have been collected in the Bresse-Dombes region.[29] Especially along the western border of the basin, in the foothills of the Beaujolais and Bourgogne Mountains, there exist numerous exposures revealing the fact that the high-level (including the so-called plateau and intermediate) gravel horizons occur at progressively lower levels, until in the region between Chalon-sur-Saône and Dijon—an area known as the Côte d'Or—they actually merge with the main Bressian (130-meter) terrace, which carries the famous *Sables de Chagny* (Delafond and Depéret, 1893, p. 213; Depéret, 1894*b*, pp. 624–625; Mayet, 1922; Depéret, Mayet, and Roman, 1923, p. 20).

This latter bed is of fluvial origin, and the orientation of its constituents is in accord with the axis of the present valleys, according to Journaux (1948, p. 69). Extensive exposures of the *Sables de Chagny* occur in the valley of the Dheune, but in the main this formation is the basal portion of a very high (130-meter) terrace, which had already been strongly and unevenly eroded by first interglacial times. As shown on the section (fig. 11), it is consequently overlain at many localities by the sands and gravels of the 30-meter terrace of the Saône, in which remains of *Elephas* (*Mammuthus*) *primigenius* and Reindeer have been recorded (Delafond and Depéret, 1893, pp. 279–284; Changarnier, 1908;

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[28] Cf. Fontannes, 1885, p. 65. Very similar torrential fan deposits, overlying Pliocene marine strata, have recently been described in the region around Figueras, just south of the Pyrenees in the extreme northeastern portion of Spain (Ribera-Faig, 1945).

[29] Depéret, 1909, pp. 140–141, where a list of the sites in question will be found; see also Delafond, 1879, p. 933; Depéret, 1885*a*, p. 256; 1885*c*, p. 122.

Depéret, Mayet, and Roman, 1923, pp. 20–21; Viret, 1926, p. 55; 1931, pp. 168–169). Although the main deposits of the *Sables de Chagny* are found on the right (west) bank of the Saône in the Côte d'Or–Beaujolais-Bourgogne region, it is also of sporadic occurrence on the left (east) bank of the river, as has been observed by Viret (1930) and Rouyer (1930).

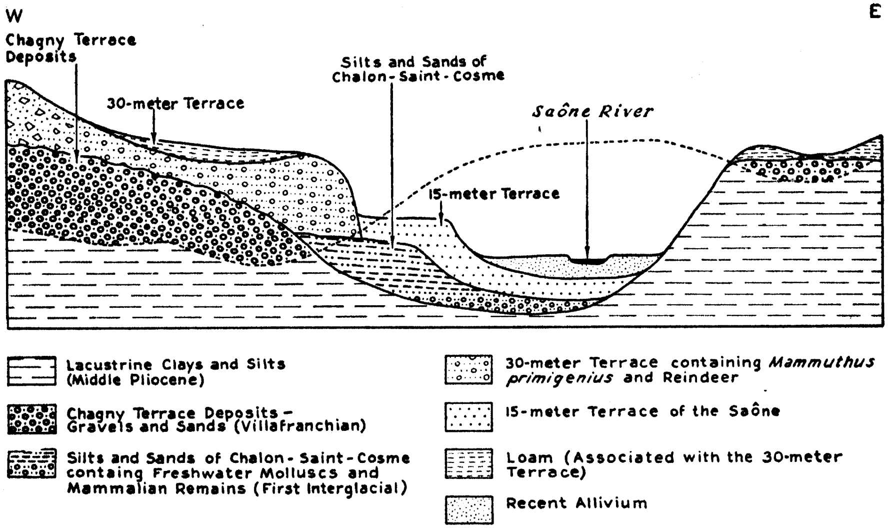
W
E
Chagny Terrace Deposits
30-meter Terrace
Silts and Sands of Chalon-Saint-Cosme
Saône River
15-meter Terrace
Lacustrine Clays and Silts (Middle Pliocene)
Chagny Terrace Deposits – Gravels and Sands (Villafranchian)
Silts and Sands of Chalon-Saint-Cosme containg Freshwater Molluscs and Mammalian Remains (First Interglacial)
30-meter Terrace containing *Mammuthus primigenius* and Reindeer
15-meter Terrace of the Saône
Loam (Associated with the 30-meter Terrace)
Recent Allivium

FIG. 11.—Diagrammatic section of the Saône Valley near Chalon, showing the stratigraphic position of the Chagny terrace deposits in relation to the silts and sands of Chalon-Saint-Cosme. (After Viret, 1931.)

An extremely rich mammalian assemblage with incontestably Villafranchian affinities is known from the sandy deposits of this main Bressian terrace in the vicinity of Chagny,[30] a town in the Saône Valley, as well as at a great number of localities north of Chagny, where rich collections of fossil bones were recovered when the railway from Beaune to Dijon was constructed during the 1880's.[31] These ferruginous sands and gravels have a very extensive distribution throughout the Bresse-Dombes Basin. Because they contain *Elephas* (*Archidiskodon*) *planifrons* Falc. (in association with the last of the mastodons), *Leptobos etruscus* Falc. and *Equus stenonis* Cocchi, as well as other forms known from the classic localities of the Asti, Val d'Arno, and Auvergne regions, there is absolutely no question concerning the paleontological affinities of this horizon. The Chagny locality provides further cognate proof that, as in the case of the Po and upper Val d'Arno Valley areas, the deposits of the Villafranchian stage are superposed

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[30] Depéret, 1885*a*, p. 261–262; 1893, p. 535; Tournouër, 1866, pp. 787–788; Delafond and Depéret, 1893, pp. 231–329; Haug, 1911, p. 1844; Mayet, Nugue, and Dareste de la Chavanne, 1920*a*, *b;* Mayet, 1922; Depéret, Mayet, and Roman, 1923, pp. 22–24; Nugue, 1924; De Prunelé, 1924; Osborn, 1942, pp. 961–963. For detailed descriptions of the main fossil locality (Bellecroix) at Chagny see Depéret and Mazeran, 1920; Depéret, Mayet, and Roman, 1923, pp. 17–22.

[31] Tardy, 1884, p. 718; Parandier, 1891; Delafond and Depéret, 1893, p. 215; Changarnier, 1908, pp. 147–150 (list of localities); Depéret and Mazeran, 1920, pp. 307–308; Depéret, Mayet, and Roman, 1923, p. 21.

directly on those of the Upper Pliocene, although the two deposits are separated from each other by a period of unknown duration, when uplift and erosion occurred. Its marine equivalent, Calabrian, is not represented in southern France,[32] where the deposits of this stage in the lower Rhône Valley and in Languedoc are all of continental facies.

The formation of the Chagny, or main Bressian, terrace and associated deposits in the Saône Basin was followed by an important phase of erosion and valley-cutting, during which the Rhône, Ain, Saône, and other rivers in the region lowered their beds 40–50 meters (Delafond and Depéret, 1893, p. 241). There ensued an aggradational stage when fine-grained sediments—silts, sands, and clays—were laid down. These are known as the Silts and Sands of Chalon-Saint-Cosme,[33] and they are found at an average height of approximately 12 meters above the level of the present streams, although they have been recorded up to a maximum height of 25 meters (Delafond and Depéret, 1893, p. 242; Depéret and Mazeran, 1920, p. 306; Viret, 1931, p. 171). That a regime in part at least of lacustrine deposition was under way in the Chalon region is indicated by the fine sections investigated by Viret (1926, pp. 55–56; 1931, p. 171), in which horizons of finely bedded, laminated clays, typical of lake deposits and recalling varved sediments, were exposed. In the vicinity of Chalon-sur-Saône (fig. 11), this so-called "Saint-Cosme terrace" has yielded remains of a typical first interglacial (Saint-Prestian–Cromer Forest-Bed) mammalian fauna,[34] including true *Bos, Megaceros hibernicus* Owen, and *Trogonterium cuvieri* Fisch. The mollusks, described by Delafond and Depéret (1893, pp. 253–256), from the clay pits at Saint-Cosme clearly demonstrate that the temperate conditions of a climate very similar to that of the region at present prevailed at the time that the sediments in question were being accumulated. The fact that in the Rhône Valley the outwash gravels and morainic debris of a major ice advance (cf. Delafond and Depéret, 1893, p. 247) directly overlie these Saint-Cosme sands and silts is further convincing evidence that the latter stratum was formed during interglacial times. In any case, the Chalon–Saint-Cosme locality is of great importance from a stratigraphic point of view; it is one of the very few sites in France where a fauna of the first interglacial stage directly overlies demonstrably Villafranchian deposits. Furthermore, just as in Italy and Sicily, an erosion interval separates the two beds.

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[32] With the possible exception of the lower valley of the Var (cf. Depéret and Caziot, 1903, pp. 322–326; Bertrand, 1904; Gignoux, 1911, p. 129; 1913, p. 336). However, marine sands and clays of the Plaisancian-Astian are extensively developed all along the coast of Provence (Lutaud, 1924, pp. 33–35). At the classic locality for vertebrate mammalian remains of the Upper Pliocene stage (Depéret's "Astian fauna"), situated near Montpellier (Hérault), the deposits have been intensively studied by the geologist Viguier (1889, pp. 390–399; 1890). Here the uppermost horizon consists of fluvial gravels, which ravine the fresh-water sands and silts of the Upper Pliocene and which have yielded typical remains of the Villafranchian elephant, *E. (Archidiskodon) meridionalis* Nesti. Furthermore, everywhere on the Roussillon plain heavily weathered sands and gravels of continental facies disconformably overlie the Upper Pliocene lacustrine series (Bourcart 1939, p. 166; 1945; pp. 44–45). It is therefore apparent that in southern France the continental Villafranchian deposits are directly superposed on those of the Upper Pliocene, although separated from them by a disconformity, just as they are in the Saône Basin, the main valley of the Rhone, the Po region, and the Arno Valley.

[33] Delafond (1886*b*, p. 80; 1890) was the first geologist to recognize this horizon as an independent stratigraphic unit in the late Cenozoic sequence of the Bresse-Dombes Basin.

[34] For lists see Adenot, 1899; Delafond and Depéret, 1893, pp. 248–253; Nugue, 1907; Haug, 1911, p. 1844; Depéret, Mayet, and Roman, 1923, p. 64.

BASIN OF PUY-EN-VELAY (HAUTE-LOIRE)

For well over a century the importance of the basin of Puy-en-Velay has been known to collectors of fossil vertebrates (cf. Robert, 1829; Aymard, 1853) as well as to students of vulcanism; but no comprehensive work dealing with the geology of the region as a whole appeared until 1892, when Marcellin Boule's famous *Description géologique du Velay* was published. This has remained the standard work on the area ever since.

The Pliocene deposits of the Puy-en-Velay region are dominantly of volcanic origin, which at one locality in the Mézenc Massif, southeast of Le Puy, have yielded a typically Pontian (Lower Pliocene) fauna (Boule, 1892, pp. 112–113). Although no vertebrate fossils have thus far been discovered in the Lower or Middle Pliocene stages in the Puy Basin proper, plant remains have been found at several localities in beds of this age. These have been intensively studied by De Saporta and Boulay (De Saporta, 1879, p. 337; Boulay, 1887; De Saporta, in Boule, 1892, pp. 116–118). The flora demonstrates that warm subtropical conditions prevailed in central France at this time. The relative poverty of Middle Pliocene[35] deposits in the Puy-en-Velay area as a whole can be explained, according to Boule (1892, p. 167), as the result of the erosion and down-cutting of the basin by the Loire River and its tributaries during the ensuing stage—the Upper Pliocene—on the basis of the stratigraphic scheme adopted in this paper.

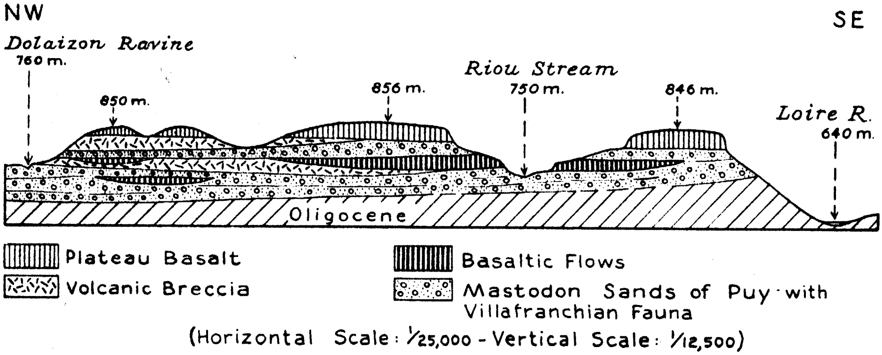

FIG. 12.—Section through the Cenozoic deposits of the Puy-en-Velay Basin between the Dolaizon ravine and the valley of the Loire. (After Boule, 1892, fig. 49.)

Following this prolonged phase of volcanic activity and erosion, an extremely important series of ferruginous-colored, alluvial deposits intercalated with volcanic material was accumulated, considerable portions of which still remain (fig. 12).[36] These fresh-water sediments, which are commonly cross-bedded and which are known collectively as the *Sables à Mastodontes du Puy,* consist in the main of fine micaceous and quartzitic sands; but gravel layers, lenses of rolled pebbles, and beds of varicolored clays also occur. At several points this bed, the surface of which is some 200 meters above the Loire, attains a total thickness of nearly 100 meters; from it a rich Villafranchian mammalian fauna has been described. Although at two localities the clay beds are prolific of diatoms and plant remains, no detailed study of them has ever been made. Boule gives the identifications of the diatoms which he collected from the clay deposits exposed near the town of Ceyssac (fig. 13), in the central portion of the basin, some 4 km. due west of Le Puy, but makes no further comments (Boule, 1892, p. 187). His list, however, was studied by Dr. Roy M. Whelden, of the staff of the Farlow Herbarium, Harvard University, who states that five of the species represented: *Melosira arenaria* Moore, *M. granulata* (Ehrenb.) Ralfs., *Navicula nobilis* (Ehrenb.) Kütz., *Stephanodiscus astraea* (Ehrenb.) Grun., and *Tetracyclus emarginatus* (Ehrenb.) W. Sm. are forms most frequently found in cold water, whereas, with respect to the assemblage as a whole, he feels that "one might safely conclude that the environment in which these diatoms grew was at least cool temperate, and more probably was decidedly cold." Dr. Whelden's observations are in complete agreement with E. Haug's statement (1911, p. 1821) that the collection of these microscopic organisms from Ceyssac reveals a very marked predominance of cold species.[37]

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[35] This corresponds to the *Pliocène inférieur* in Boule's stratigraphic scheme (cf. Boule, 1892; 1893*a, c*).

[36] For a detailed description of this interesting series see Boule, 1892, pp. 168–201; see also Chaput, 1917, pp. 59–60; 1919, pp. 85–86; Baulig, 1928, p. 219.

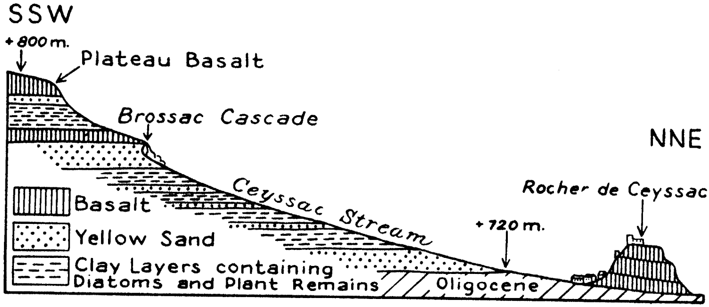

FIG. 13.—Geologic section showing the clay beds of Villafranchian age exposed near Ceyssac in the vicinity of Le Puy (Haute-Loire). (After Boule, 1892, fig. 49.)

In addition to diatoms, the Ceyssac clay beds have yielded a rich collection of fossil plant remains.[38] As De Sapotra and Depape have pointed out, in comparison with the older floras of the Massif Central, on the one hand, and those of the Rhône and Ain valleys, on the other, the Ceyssac flora offers a very striking contrast; for, in it, all the tropical to subtropical forms (*Bambusa*, *Magnolia*, *Sabal*, *Zelkovia*, *Sassafras*, etc.) have completely disappeared; indeed, only a very small number of forms reported from any of the demonstrably earlier localities has survived. The Ceyssac flora, which included *Pinus*, *Picea*, *Abies*, *Vaccinium*, *Carpinus*, *Acer*, *Alnus*, *Ulmus*, *Populus*, *Salix*, *Fraxinus*, *Crataegus*, and *Pyrus*, indicates that during Villafranchian times the climate was somewhat wetter than it is in this region at present and, furthermore, that the mean annual temperature of the Puy-en-Velay Basin was approximately 10° C. (50° F.),

[37] See also Lauby (1910, pp. 293–294), who mentions the interesting fact that certain "cold" types of fresh-water mollusks are fairly abundant at the Ceyssac locality. In this connection a study of the insect remains that occur in the Ceyssac clay beds (Boule, 1892, p. 185) would doubtlessly prove interesting.

[38] De Saporta, 1873, pp. 226–230; 1879, pp. 344–345; also Boule, 1892, pp. 185–187; Depéret, 1893, p. 527; Lauby, 1910, p. 293; Laurent and Marty, 1927, pp. 35–36; Depape, 1928, pp. 45 and 66–67.

or 2° C. cooler than the present mean (12° C. = 53.6 F.) for this section of France (see Depape, 1928, p. 81). Therefore, an analysis of the flora is consistent in every respect with that derived from a study of the diatoms, although in neither case have the results been worked out on a numerical basis, according to the percentage frequencies of the various forms represented in terms of the total assemblage. The climatic interpretation based on the above is significant, however, because it implies that at the time of the accumulation of the Mastodon Sands of Puy-en-Velay the mean annual temperature in this basin was not only considerably colder (possibly as much as 8° C., or 14.4 F. *lower*) than it has been previously during the Pliocene, but also it was some 2° C. (3.6 F.) colder than it is on the average in this region at present. This evidence is in complete accord with the view expressed elsewhere, namely, that the Villafranchian in southern and western Europe represents a general phase when continental conditions prevailed,[39] when the sea was at a relatively low level, and when a glacial episode was already making itself manifest.

The fossil vertebrates known from the Mastodon Sands of Puy[40] were long regarded as demonstrating a Middle Pliocene age for this formation, but, because the animals represented constitute an assemblage that is incontestably Villafranchian (see Depéret, 1885*a*, p. 255; 1893, p. 536; Haug, 1911, p. 1821; Depéret, Mayet, and Roman, 1923, p. 34), this view is no longer tenable. Indeed, the fauna from the deposits in question is very similar in all its essential elements to those recorded from the classic sites of Perrier, Senèze, Chagny, the upper Arno and the Asti region (Depéret, 1893, p. 524; 1909). In spite of the fact that this fauna is known from a number of localities in the Puy Basin, it is surprising that no data have ever been published concerning the relative stratigraphy of the remains in question. As the *Sables à Mastodontes du Puy* are up to 100 meters thick, it is very probable that they cover the entire range of the Villafranchian, although they may have been laid down relatively rapidly in a crater lake. Until the Velay Basin is resurveyed and studied by a geologist in company with a paleontologist, however, there seems to be absolutely no justification for the contention that the Villafranchian stage can be subdivided into "early" and "late" substages on the basis of the evidence from this region.

In the Puy Basin, as in the case of the Saône Valley and the Italian peninsula, the Villafranchian stage was closed by uplift and erosion. According to Boule, it is probable that epeirogenic movements related to a new cycle of vulcanism occurred at this time, a view which is supported by the marked disconformity between the Mastodon Sands and the overlying volcanic flows, which is observable at many localities (Boule, 1889, p. 278; 1890, p. 948; 1892, pp. 200–202; 1893*d*, p. 518). The age of these eruptions is established on the basis of the paleontological evidence from La Malouteyre, on the northeastern flank of Mount Denise,[41] and Sainzelles, a small village situated some 2½ km. northwest of Mount Denise. At La Malouteyre[42] the post-Villafranchian deposits are mainly of volcanic origin, and they have yielded a fauna which is definitely of first interglacial (Saint-Prestian) age.[43]

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[39] As pointed out on p. 399, the sea had completely withdrawn from the Rhône Valley by this time; Italy was likewise slowly emerging in a general north-to-south direction (see p. 390).

[40] For lists and descriptions see: Depéret, 1885*a*, p. 255; 1885*b*, p. 121; 1893, p. 536; 1907, p. 425; Boule, 1889, p. 275; 1892, p. 188; 1893*b*, p. 510; 1907; Laurent, 1907, p. 388; Haug, 1911, p. 1821; Depéret, Mayet, and Roman, 1923, pp. 35–37; Viret, 1942.

[41] The extinct volcanic cone of Mount Denise is located approximately 3 km. northwest of Le Puy.

The fossil locality of Sainzelles (fig. 14), which was discovered nearly a hundred years ago (Aymard, 1853), consists of re-worked volcanic tuffs, apparently accumulated during and immediately subsequent to the erosion interval previously mentioned. These deposits are overlain by an extensive sheet of basalt; they may be easily distinguished from the Mastodon Sands by their grayish tint, by the large size of their constituent pebbles, and by the angular nature of the gravels which they contain. As Depéret originally pointed out,[44] the Sainzelles fauna is very close to that of Saint-Prest, near Chartres (Eure-et-Loire), on the one hand, and to the assemblage from Chalon–Saint-Cosme in the Saône Valley (see p. 399), on the other. Furthermore, at the latter locality, as well as in the Puy Basin, the beds yielding this first interglacial fauna occupy an analogous stratigraphic position with respect to the underlying deposits of the Villafranchian stage. It is to be regretted, however, that no plant remains have thus far been discovered at the Sainzelles–La Malouteyre localities of the Puy region, because it would be interesting to compare them with the earlier Villafranchian assemblage from Ceyssac. Nevertheless, as Haug (1911, p. 1822) has pointed out, the presence of Hippopotamus (*H. major* Cuv.) is very convincing evidence that, at the time that this bed was accumulating, a relatively warm interglacial climate prevailed in the region.

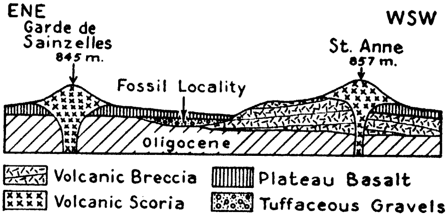

FIG. 14.—Section showing the stratigraphic position of the fossiliferous gravels at Sainzelles in the Puy-en-Velay Basin (Haute-Loire). (After Boule, 1892, fig. 60.)

CONCLUSIONS

The stratigraphic facts briefly outlined above are summarized in the table of figure 8; their significance with regard to the problem of the boundary between the Pliocene and the Pleistocene is at once apparent. Collectively they demonstrate the following:

1. That the deposits containing the classic Villafranchian fauna, considered as representing a Villafranchian stage, not only consistently overlie the Upper Pliocene in Europe but are also, in turn, immediately overlain by deposits that clearly belong to the first interglacial (Cromer Forest-Bed–Saint-Prestian) stage.
2. That, with the exception of peninsu-

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[42] For references to the La Malouteyre fauna see Boule, 1889, p. 278; 1892, p. 202; 1893*d*, p. 518; Déperet, 1893, p. 536; 1909, p. 142; Haug, 1911, p. 1822; Depéret, Mayet, and Roman, 1923, pp. 38, 151.

[43] The fossil locality of Saint-Prest, situated in the valley of the Eure, 8 km. northeast of Chartres (Eure-et-Loire), has been described in detail by Denizot (1927; see also Haug, 1911, p. 1807; Depéret, Mayet, and Roman, 1923, pp. 60–62; Joleaud and Alimen, 1945, p. 95). The Saint-Prestian stage, which is the equivalent in France of the Cromer Forest-Bed horizon of southeastern England, was considered to be *Pliocène supérieur* by Boule (cf., e.g., Boule, 1892, p. 201, 1946, p. 165).

[44] Depéret, 1885*a*, p. 263; also Joleaud and Alimen, 1945, pp. 15 and 144. For lists of the Sainzelles fauna see Depéret, 1885*a*, p. 263; 1893, p. 536; 1909, p. 142; Boule, 1892, pp. 203–206 and 216; 1893*d*, p. 523; 1893*e*, p. 88; Haug, 1911, p. 1822; Depéret, Mayet, and Roman, 1923, pp. 62–63.

lar Italy, where a marine regime still prevailed, the deposits in question, which are of continental facies (relatively coarse fluvial and fluvio-lacustrine sands and gravels) represent a very marked break in the sedimentary cycle that had obtained in southern and southwestern Europe throughout the Middle and Upper Pliocene.

3. That during this part of the late Cenozoic period a marked lowering of the temperature occurred, which is consistent with the view that the Villafranchian deposits were laid down under the conditions of a pronounced climatic deterioration, which one would expect to accompany the onset of the first glacial stage.
4. That immediately prior to the initiation of this phase of climatic change a series of orogenic movements, which resulted in uplift and erosion, profoundly modified the entire Apennine tract. Furthermore, contemporary coastal deposits in Italy and Sicily prove that the Calabrian Sea was of a transgressive nature and that between the silts, sands, and clays of this stage and those of the Plaisancian-Astian series an important marine regression occurred.

With regard to the stratigraphic evidence as a whole, the sequence in southern and southwestern Europe, therefore, appears to be complete from the base of the Pliocene through to the end of the first interglacial stage. Each of the five major subdivisions (Pontian, Middle and Upper Pliocene, first glacial, and first interglacial stages) is represented by a series of geological horizons. Because there seems to be no doubt whatsoever concerning either the first interglacial dating of the Saint-Cosme–Sainzelles-Sicilian horizons or the pre-Pleistocene (i.e., Upper Pliocene) age of the Plaisancian-Astian and equivalent deposits, the Villafranchian beds should have been laid down during the part of late Cenozoic time separating these two latter stages. In other words, the stratigraphic evidence clearly indicates that the Villafranchian sediments of southern and southwestern Europe were accumulated during the first major glacial episode of late Cenozoic time. On this basis the Villafranchian is regarded by definition as marking the base of the Pleistocene rather than as representing the final stage of the Pliocene.

In view of the fact that the primary concept on which the science of geology is based is the Law of Superposition, one actually encounters considerable difficulty in retaining the Villafranchian stage in the Pliocene epoch; for, by so doing, the first glacial stage is represented stratigraphically only by a lacuna, with *no* deposits as yet reported from anywhere throughout the vast region under consideration which are referable to it. That such an event, which reflects a major climatic change of the first order of magnitude, should not have been widely recorded seems very hard to understand. Indeed, it would be difficult to attribute the huge torrential outwash fan deposits in the western Alpine forelands as due to any cause other than the relatively sudden impact of a glacial climate on the depositional processes operating in the region. With respect to the western portion of the Po Plain, this concept was originally propounded by Sacco (1890, p. 331) almost sixty years ago. In any case, the fact that a marked lowering of the temperature occurred at this time is proved by the following critical evidence: (*a*) the invasion of the Mediterranean Basin by "cold" boreal types of mollusks and Foraminifera, which represent up to

10 per cent of the total marine invertebrate fauna; (*b*) the plant remains from (1) Lodi in the northwestern section of the Po Plain, (2) various localities in the upper Val d'Arno where Zone b of the upper series of the basin sediments has been investigated, and (3) the Villafranchian clay beds exposed near Ceyssac in the Puy-en-Velay Basin of Haute-Loire; (*c*) the diatoms from the latter locality; and (*d*) the congeliturbated layers that have been recently described in the outwash fans of Villafranchian age in the Dijon region. From a chronologic point of view this series of events seems to be closely related and to be the result of the same general climatic change which led to the temporary establishment of a regime during which conditions were both cooler and wetter throughout southern and southwestern Europe than they are at present.

Immediately prior to the initiation of the Villafranchian-Calabrian stage in Italy, an important phase of orogenic movements occurred in the Apennines, which caused the Plaisancian-Astian beds to be uplifted and tilted and the formation of an erosion surface on those portions that were exposed. At the same time and apparently directly related to the mountain-building processes taking place in the Apennines, there was an abrupt break in the cycle of marine sedimentation which had been under way in the Mediterranean since the close of the Pontian. Although the evidence for this break is not everywhere apparent in the Messina region and Calabria, many localities elsewhere in Sicily, southern Italy, and along the Po Plain clearly demonstrate the transgressive nature of the Calabrian. This followed a phase of marine regression that marks the close of the Plaisancian-Astian series. In comparison with the total span of Pliocene and Pleistocene time, however, the interval represented by these latter events was presumably of short duration. Finally, the fact that Europe was invaded at this time by new forms of animals is regarded as significant, in view of the fact that, normally, large migrations of terrestial mammals are directly related to broad geographical changes.

This problem is vast, and it opens up intriguing fields for further studies. Among the projects that should be given high priorities in this connection are the following:

1. An investigation of the dates of the "late Pliocene" epeirogenic movements in the Alpine region, references to which in the literature are at present somewhat confusing in most cases and not clearly defined stratigraphically.
2. Further soil studies as a means of establishing the climatic conditions of the time of formation of these early deposits, as well as the processes by which they were formed.
3. A reinvestigation of the Ceyssac plant-bearing and diatomaceous deposit in the Puy-en-Velay Basin by modern paleobotanical methods.
4. A percentage frequency determination of the Calabrian and Sicilian molluscan faunas from localities selected to represent northern as well as southern Italy and Sicily.

It is further suggested that the actual beds from which the mammalian fossils have been collected in the past be carefully rechecked in the light of modern field methods. With the sole exceptions of the Senèze, Roca Neyra (Perrier), and Bellecroix (Chagny) localities, the paleontological material in general is rather vaguely grouped and referred to, both with regard to locality and stratum. Apparently, the main objective of the early paleontologists was to amass as great a bulk as possible of fossil bones of various types, which may account for the fact that remains of small carnivores, rodents, insectivores, etc., are either very rare or supposedly nonexistent. The presence of these small forms would furnish critical climatic evidence in many cases. On the other hand, there is an abundance of large animals—proboscidians, rhinoceroses, Bovidae, horses, etc.—which are of little help for environmental studies. In only a very few cases, however, is the precise stratigraphic horizon of a given find indicated. As the deposits in the upper Val d'Arno and in the Puy-en-Velay Basin are extremely thick, they presumably represent considerable time. It is reasonable, therefore, to expect that some phylogenetic change in the fauna from base to top probably took place. In view of the fact that during the time represented by the stratigraphic break at the base of the Pleistocene many of the higher vertebrates seem to have been passing through an extremely critical evolutionary stage, this problem would almost certainly repay careful investigation. In the meantime the present data are not regarded as sufficient to permit subdivision of the Villafranchian stage, the beginning of which is believed to coincide with that of the Pleistocene epoch.

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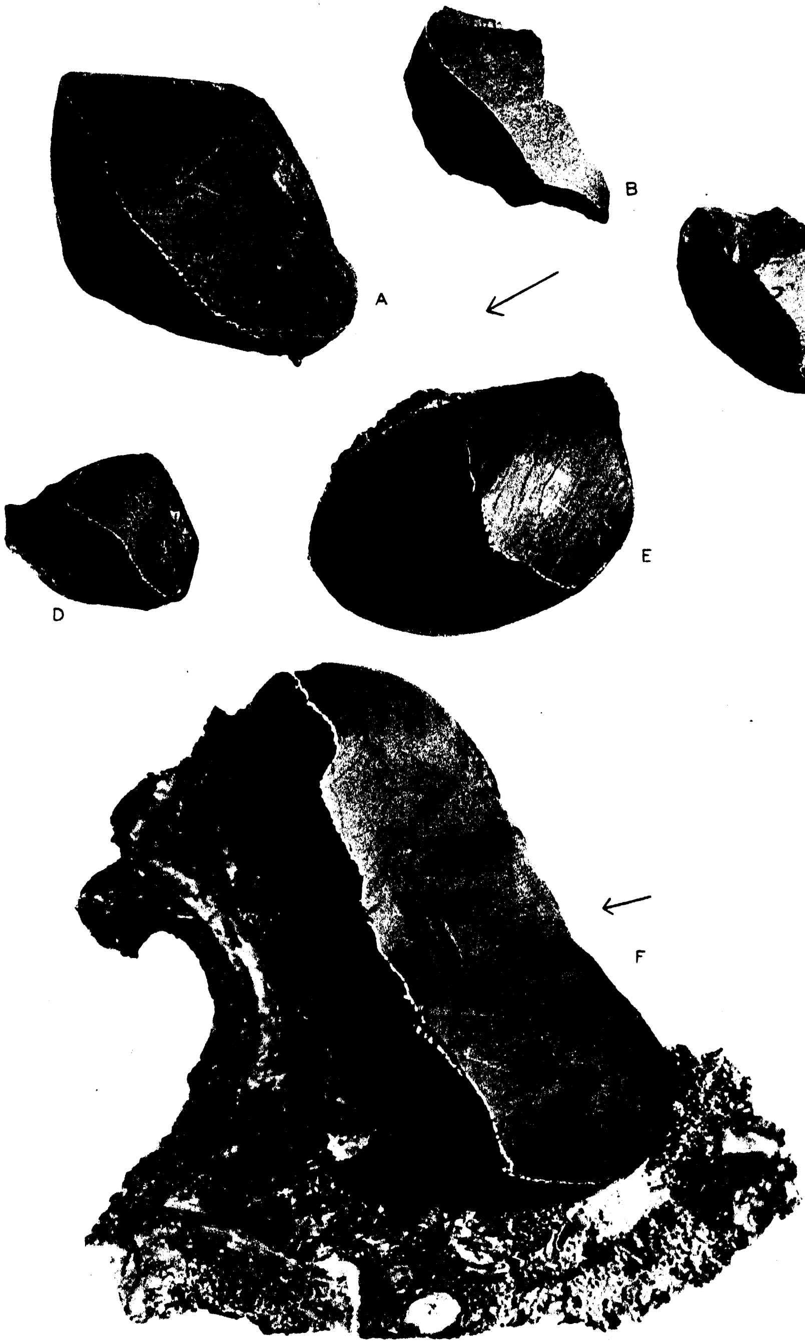

Wind-faceted pebbles

GEOLOGICAL NOTES

SOME REMARKABLE WIND-FACETED PEBBLES FROM NEAR THE MOUTH OF THE UMHLANGAKULU RIVER, SOUTHERN NATAL, SOUTH AFRICA[1]

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In years past I have made several field studies upon the cutting of facets on pebbles by the action of wind-driven sand, and particularly upon the number and direction of facets in relation to the dominant wind directions.

The approach has been from several angles. First, to evaluate the effect of original pebble shape, studies were made of angular material shed from a limestone outcrop upon the beach near the Ure River, Marlborough, New Zealand (King, 1936*a*). Here only two opposite wind directions along the beach could be effective for sand-blasting. Material was studied at all stages from freshly shed fragments to superbly faceted dreikanter and pyramidalgeschiebe. The conclusion was reached that, *with angular material*, original shape exercised considerable control over the final form assumed under sand-blasting.

Later, deposits near the mouth of the Clarence River in the same province (King, 1936*b*) (where, again, only two opposed wind directions could prove effective) afforded opportunity for the study of the relations of facets to dominant wind directions. Here river-worn, rounded boulders lay, abundantly faceted, on the raised beach. The conclusion reached was that *upon originally rounded materials* wind direction controls the orientation of facets.

In these and other cases (King, 1949), a major difficulty has been to demonstrate that the pebbles lay in relatively undisturbed condition and that they had undergone no rotation during their formation in either a horizontal or a vertical plane. Pebbles and facets were always recorded as completely as possible (King, 1936*b*); but, even so, the degree to which disturbance might enter, especially owing to the trampling of man and other animals, remained largely uncertain.

Wind-faceted lydianite pebbles projecting from a fossiliferous marine conglomerate of Cretaceous age just south of the Umhlangakulu River, southern Natal, are therefore of special interest. Here the conglomerate crops out along the present-day beach in such a manner that many of the pebbles, held fast by their nether surfaces in the Cretaceous matrix, have had their upper parts faceted by wind-driven sand. They cannot have been rotated at any time; study of their facet

PLATE 1

A–E, Wind-faceted stones, all oriented as found and with wind direction shown by arrow, from the beach south of Umhlangakulu River mouth, Natal, South Africa. The pebbles projected from the surface of a Cretaceous conglomerate in which they were embedded, fragments of which still adhere to them. Faceting is by wind-driven sand of the present beach. All facets are nearly at right angles to the only effective wind direction.

F, Crudely faceted pebble *in situ* in fossiliferous Cretaceous conglomerate. The faceting by wind is of recent date. Beach south of Umhlangakulu River mouth, Natal, South Africa.

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[1] Manuscript received October 16, 1948.

directions may be expected to bear more than usual significance.

That the faceting is not of Cretaceous age but is of modern origin is clear because (*a*) none of the other pebbles in the body of the conglomerate bear any facets but, instead, are rounded and water-worn, and (*b*) none of the pebbles exposed superficially on the adjacent, wave-swept, tidal rock platform bear any facets. The faceting, let us repeat, is a product of the present situation on the existing wind-swept beach.

Only winds blowing from the southwest to the northeast, parallel with the trend of the beach, are effective because the landward side is cut off by high dunes, fixed by vegetation, and the winds off the sea lack the long fetch of sand necessary to faceting. Winds blowing from northeast to southwest are ineffective because the lagoonal outlet of the river traps any pronounced drift of sand from that direction.

The controls are thus stringent: the pebbles are held fast in matrix and are incapable of rotation, and there is but a single direction of effective sand-blast.

Unfortunately, the number of faceted pebbles is small, barely a score, though many more doubtless lie buried beneath the beach sand farther down the coast. Nevertheless, the facets are distinct and distinctive—no more than one to each faceted pebble, and every facet faces directly to the southwest, i.e., upwind. This is so whatever the orientation of the longer axes of the pebbles, so that in some cases the facets are placed laterally and in other cases "on the nose" of the pebbles. Some are intermediate in position.

Thus it appears that, on originally rounded material, effective wind direction alone is responsible for the direction in which facets are cut, also that the normal direction is at right angles to the wind. This conclusion is in harmony with the census of wind-faceted pebbles from the Clarence River mouth (King, 1936*b*) and the experiments of Schoewe (1932).

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GEOCHEMICAL RESEARCHES IN JAPAN[1]

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Geochemical research in Japan began in 1913 when Y. Shibata spectrochemically analyzed naegite, a variety of zircon containing rare earths, thorium, and uranium. Shibata continued the research on Japanese minerals containing rare elements with collaborators and published the results in 1923. This was the first report on geochemical research in Japan, although the word "geochemistry" was not used.[2] These investigations were followed by those of Kimura, and since that time about forty reports have been published. Besides Kimura, S. Iimori and J. Takubo also studied the minerals containing rare elements. Iimori in particular studied the minerals found in Korea. The occurrences of the following minerals and

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[1] Manuscript received October 27, 1948.

[2] The Japanese word *Tikyu Kagaku*, corresponding to "geochemistry" and meaning literally the "chemistry of the earth," was invented and used by Y. Shibata in 1926 for the first time.

the results of their analyses were described: allanite, autunite, beryl, calciogadolinite, columbite, enalite, euxenite, fergusonite, gadolinite, ilmenorutile, ishikawaite, pitchblende, samarskite, tantalite, thorogummite, torbernite, uraninite, xenotime, yttrialite, zircon.

The minor constituents of Japanese shales were studied by Minami in Göttingen. Volcanic rocks of volcanoes Asama, Izu-Oshima, etc., were analyzed by I. Iwasaki, and the mean composition of Japanese igneous rocks as determined by him is given in table 1. M. Ikawa studied plutonic rocks from the Dando District of central Japan, which resemble in geological and petrological properties the trondhjemites of Norway. Trace elements in Japanese soils, such as arsenic, iodine, and boron, were determined by K. Hirai. Recently, a very interesting aqueous inclusion was found in nepheline basalt in Iwami-Nagahama. This is thought to be the residual water from magmatic differentiation. There is a close resemblance between the chemical composition of this water and that of sea water.

The hydrosphere is better known than the lithosphere. The geochemical studies of hot springs in Japan were made by Kimura, Kuroda, Nakai, Okuno, Uzumasa, and others. Kimura and Kuroda analyzed more than three hundred spring waters spectrochemically and found the following trace elements: Ag, As, Au, B, Ba, Be, Bi, Co, Cs, Cu, Cr, Ga, Ge, Li, Mn, Mo, Ni, P, Pb, Sb, Sn, Sr, Ti, V, W, Zn, and Zr. The fluorine content of mineral springs was determined by Okuno, and their radioactivity was studied by Nakai and Kuroda. The heavy-water content of many kinds of natural waters was determined by Oana.

Chemical studies of Japanese lakes were made by the late S. Yoshimura in 1926. K. Sugawara also made a thorough study of lake metabolism. In 1928 I. Okada made a geochemical study of Dabsnor, a soda lake in Manchuria, and discovered gaylussite ($Na_2CO_3 \cdot CaCO_3 \cdot 5H_2O$) for the first time in the Orient. Extensive studies of brine from Szechwan Province in China were made by the late S. Goda in Shanghai. The chemical composition of river waters was studied by E. Kurashige and Y. Miyake.

Studies of the sea include the work of Y. Miyake, who analyzed the oceanic salts of the western Pacific Ocean and determined the vertical distribution of minor constituents in the Kurosio region. The radium content of the Yellow Sea and the China Sea was determined by Goda in 1930, and his results ($0.04 \sim 0.15 \times 10^{-12}$ gm radium/l) are in good agreement with the data of R. D. Evans. The radium content of ocean-bottom sediments of the western Pacific Ocean collected by the naval sounding ships was determined by H. Hamaguti. M. Ishibashi determined the quantities of trace elements, such as Au, Cs, Cu, Li, Pb, and Ra, in sea water.

TABLE 1

MEAN COMPOSITION OF JAPANESE IGNEOUS ROCKS

Oxide	Per Cent	Oxide	Per Cent
SiO_2	59.84	CaO	6.54
Al_2O_3	16.84	MgO	2.92
TiO_2	0.74	Na_2O	3.00
Fe_2O_3	2.80	K_2O	1.49
FeO	4.55	H_2O	1.28
MnO	0.21	P_2O_5	0.22

The study of the natural gases in Japan was first made in 1923 by N. Yamada in a search for helium. The highest helium content found was 0.3 per cent, but the quantity discharged was very small. The helium content of monazite, beryl, and other minerals was determined by J. Sasaki and B. Yamaguchi. Gases from fumaroles and mineral springs were studied by Noguti. The gaseous impurities contained in the air, such as ammonia, sulphur dioxide, and the oxides of nitrogen, were studied by Y. Miyake, and the seasonal variations of their contents were also determined.

Geochemical researches in Japan were influenced equally by European and American publications in geochemistry. *The Data of Geochemistry* of Clarke and reports of Goldschmidt are well known in Japan, and the *Géochimie* of Vernadsky was translated into Japanese in 1933.

Geochemical research in Japan has been prosecuted mainly by chemists rather than geologists or mineralogists, and this is probably one of the reasons why the lithosphere is less investigated than the hydrosphere.

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A FLORIDA LANDSLIDE[1]

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During the first week in April, 1948, a landslide carried about 4 acres of a farm owned by W. D. Pitts of Greensboro, Florida, into an adjacent valley. The magnitude of the landslide, as well as certain other of its features, makes it worthy of note.

Greensboro is situated in northwestern Florida on a low plateau about 290 feet above sea level. The plateau is underlain by horizontal strata, probably of Miocene age. Regional youth is shown by flat-topped interfluves, but the larger valleys are mature or old. Because the larger valleys are cut less than several hundred feet into the partially consolidated sediments, the available relief is less than the critical value (Glock, 1932). It follows, therefore, that the interfluves may well go from youth to old age without divide maturity through lateral planation by the larger streams.

The landslide moved about 4 acres of land lying on the gentle slope (not exceeding 10°) of a wide spur extending into a swampy east-west valley. The ground had been thoroughly saturated by a 16-inch rainfall during the 30 days preceding the slide. This rain marked the culmination of a record 12-month fall, so that the region was in flood condition. For the 12-month period ending March 31, 1948, 89 inches of rain had fallen, against an average of 55 inches.[2]

Approximately 146,000 cubic yards of reddish, partially indurated clayey sands of the Hawthorne formation broke away from the hillside, leaving a horseshoe-shaped scar and a cliff of similar shape about 45 feet in height (pl. 1). The material funneled through the narrow opening of the horseshoe over a low lip into the marshy valley bottom. Because of the damming effect of the dense forest, the material did not cross the valley but spread out up and down the valley for several hundred yards, much as wet cement flows when poured from a mixer onto a flat surface. The floor of the scar is almost on a level with the valley bottom and shows very little slope.

The land composing the upper part of the

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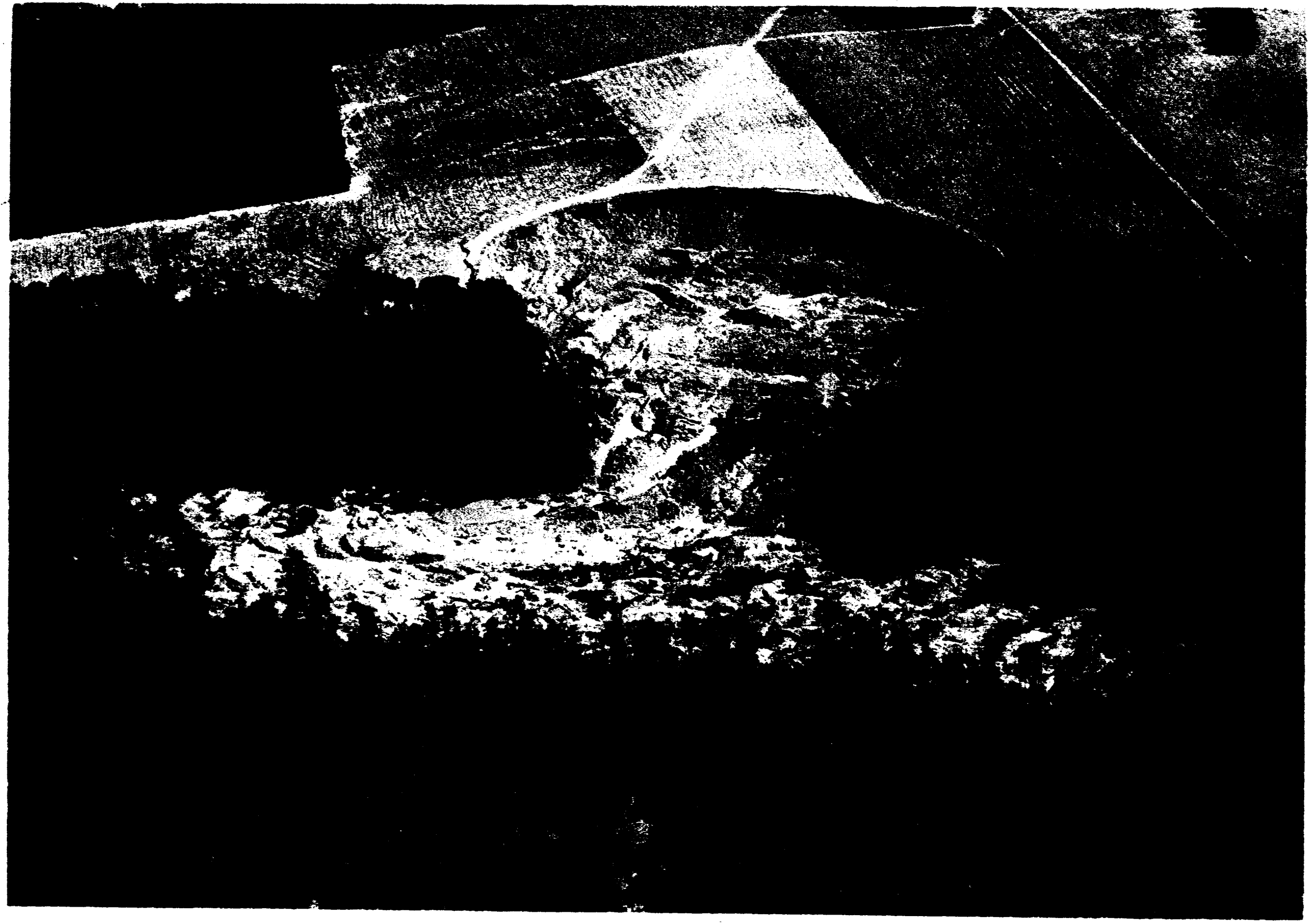
[1] Manuscript received August 30, 1948.

[2] The rainfall for the 3 months of 1948 was 23 inches, against an average fall of 12.50 inches for these months.



Aerial view of slide, from the north

landslide was a cultivated field, whereas the lower part was forested with tall pines and other trees. The swampy valley was likewise heavily wooded. The flow demolished and buried the wooded area of the swamp and overturned the large pines on the hillside. Trunks and other parts of the trees may be observed at all angles in some parts of the debris, although most of the forest was completely buried.

According to Sharpe's classification (1938), the Greensboro landslide is primarily an earthflow of the more rapid type, with slumping around the upper margins. The general shape, including the bottleneck, is somewhat comparable to the 1898 flow near St. Thuribe, Quebec, although in rate of movement it is assumed that the Greensboro flow was more rapid (Sharpe, p. 51). There was no eyewitness, but the general appearance indicates completion of movement in minutes rather than in hours.

The sediments involved are not completely indurated, and there seems to be no definite indication of a clayey or otherwise slippery plane surface on which the material slid as a whole. However, the Hawthorne formation has many lenses of fairly pure clay, which would greatly facilitate the flowage of saturated sediments. Furthermore, the sand of the formation is commonly admixed with varying amounts of clay.

After the flooded streams in the vicinity subsided, several small ponds fed by springs remained at the bottom of the scar, but no springs appeared in the walls of the cliff. Thus it seems probable that the water table at the time of the flow had been raised high enough to saturate thoroughly the otherwise comparatively well-drained layers forming the bottom of the flow. The flowage of this saturated zone may well have instigated the landslide.

Echelon mounding of the landslide topography indicates that the mass did not move as a whole but that the lower parts broke away as an earthflow, to be followed immediately and in turn by higher segments, each segment slumping toward the center of the amphitheater formed and thus enlarging its horseshoe shape. This shape is more easily accounted for by such movements than by a simultaneous slipping of the whole mass.

When all sections of the land had broken away and were well on their way to the valley, the early deposits began to dam the flowage of debris in the rear. This is indicated by the general aspect of the floor of the amphitheater, the floor being somewhat lower near the cliff than in the middle of the bowl and sloping gently from this point toward the valley floor.

Probably the most unusual feature of the whole flow is the low angle of the slope of the land previous to the slide, the maximum slope not exceeding 10°. Thus it is evident that the floor of the amphitheater, 45 feet below the top of the cliff, slopes very gently toward the valley and that the material which moved on the bottom moved on a very slight slope indeed.

Considering the high annual rainfall of this section of Florida and the semiconsolidated nature of the sediments, it might be reasoned that flows of this type and magnitude would not be unusual. A few somewhat similar small landslides more closely resembling slumps and at least one large incipient landslide in which only the cracks have formed have been noted in Florida this year. However, no actual landslip has been seen which is comparable in magnitude to that at Greensboro.

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DISCUSSION

THE DISTRIBUTION OF OXYGEN IN THE LITHOSPHERE AND OXYGEN IN ROCKS: A BASIS FOR PETROGRAPHIC CALCULATIONS

A DISCUSSION[1]

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Tom F. W. Barth (1948*a*, *b*) has discussed the distribution of oxygen in the lithosphere and outlined a new method of petrographic calculation. The latter subject was also discussed by Barth in another paper (1947).

Barth's paper on the distribution of oxygen in the lithosphere contains a restatement of what Barth calls "Ramberg's principle." "Ramberg's principle" has been discussed in another paper (1947), where I raised objections against the applicability. The chief objection is that the chemical activity of a crystalline substance is dependent not only on the temperature and pressure but also on the surrounding phases.

It is well known to all chemists working with reactions in the solid state that traces of alien ions in a crystal lattice have considerable influence on the activity of the crystal. It is generally unknown how the solubilities of different substances in different crystal lattices are related to pressure. Nothing can be stated, therefore, as to the variation of the chemical potential of a mineral component with pressure. "Ramberg's principle" is valid only in a pure system in a vacuum. In the earth the reactions may develop just opposite to the direction to be expected according to this principle.

Barth's calculations of the stability ranges of the different iron oxides would therefore, at best, be valid only for a globe consisting solely of oxygen and iron and cannot be directly applied when conditions in the earth are discussed.

At this point it is worth mentioning that the transition: $4Fe_2O_3 + Fe^{2+} = 3Fe_3O_4 - 2e$ involves a considerable increase in volume as follows:

4 hematites represent a cell volume = 204 Å^3,
3 magnetites represent a cell volume = 225 Å^3.

It can hardly be correct to calculate the distribution of iron oxides in a gravitational field without taking this fact into consideration. Barth's assumption might easily come in opposition to the Le Chatelier principle.

That an increase in pressure actually can break chemical bonds is well known to chemists. Brown and Patterson (1948) have dealt with a somewhat similar problem. They are, however, very careful in their conclusions and state (p. 92) that "although existing data, such as shown in figure 2, support such an assumption, definite proof must await refined thermochemical calculations, more precise measurements of distribution coefficients, and broader theoretical foundation of the effect of high pressure on partial molar volumes."

Barth knows that he is dealing with rather hypothetical and idealized conditions. In the last half of his paper, therefore, he deals with the applicability of his calculations. He finds that in some cases, e.g., the distribution of uranium, chemical forces make "Ramberg's principle" invalid. In the case of oxygen, however, he does not think that the chemical affinities are of the same importance. On the basis of oxygen content of olivine, basalts, "average igneous rock," and "ichor granite," he concludes that the oxygen in the lithosphere shows a tendency to be distributed according to to "Ramberg's principle."

If his table 1 (p. 42) is more closely considered, it is seen that the oxygen content rises about 1.1 per cent in going from basalts to "ichor granite"; at the same time the silicon content rises 31.5 per cent. As quartz contains almost 99 per cent oxygen by volume, an increase in the quartz (and feldspar) content of a rock will naturally involve an increase of the relative oxygen content. As the rocks rich in quartz (and feldspar) are lighter than the average lithosphere, it is not necessary to assume any diffusion process or any "Ramberg's principle" to explain the fact that the rocks of the upper sial are richer in oxygen than are the rocks of the deeper strata. It is not necessary to assume any diffusion or any "Ramberg's principle" to explain why the upper part of a milk bottle is richer in fat than the lower part.

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[1] Manuscript received September 29, 1948.

As stated by Bowen (1948), Niggli and Brandenberger (1948), and myself (1947), the diffusion coefficients of ions in solid mineral lattices are far too low to cause any transport of matter worth mentioning if the distance is of the order of 10–100 km. Even diffusion coefficients as high as 10^{-4} cm²/sec (as in water) are too low to permit diffusion of any importance in one million years. The diffusion coefficient in mineral lattices are by no means as high as 10^{-4} cm²/sec even at an elevated temperature. Most probably the right value in the deep, but still crystalline, part of the lithosphere is about 10^{-15} cm²/sec. Bowen (1948) gives diffusion coefficients of about 10^{-10} at 1,400° C. in rock-forming minerals. In uraninites I have found (1949) coefficients of $2\text{–}4 \cdot 10^{-22}$, in uranothorites $1 \cdot 10^{-18}$, and in samarskites $4 \cdot 10^{-19}$. These values are the average diffusion coefficients of lead since the crystallization of the minerals.

For example, we may assume that the diffusion coefficient of an element in the rock-forming minerals is as high as 10^{-12} cm²/sec (a value probably very much too high). Let the rock in question contain 0.1 gm. of the element in question per cubic centimeter, and let the activity of the element be doubled by raising the pressure 300 atm. (equivalent to 1,000 meters of overlying rock). (The pressure increase necessary to double the activity will probably by much higher.) These assumptions give an active gradient of the order of 10^{-6} gm/cm⁴. If the diffusion coefficient is 10^{-12} cm²/sec, 10^{-18} of the element in question will diffuse through a cross-section of 1 cm² in 1 sec., i.e., 0.03 mg. in a million years. Such a diffusion is without any geological significance.

Barth's table 1 (p. 42), showing the distribution of oxygen in the lithosphere with various depths, is a statement of fact. It is not, however, related to the degree of oxidation of the iron, as might be concluded from Barth's paper.

According to Clarke and Goldschmidt (course in petrology at Oslo University), the FeO/Fe_2O_3 ratios in weight percentage of different rocks are as shown in the accompanying tabulation. These values are all higher than the value for magnetite (0.45). It is obvious that the escaping tendency of oxygen in a rock is dependent upon how the oxygen is chemically combined, e.g., the oxidation state of the iron. The volume percentage of oxygen in a rock is an indicator of the escaping tendency, *only as long as all the cations occur in the same amount*. Thus the escaping tendency of oxygen in quartz with about 99 per cent by volume of oxygen is not necessarily greater than in K_2O with about 30 per cent by volume of oxygen. A mere statement of the volume percentage of oxygen in rocks at various depths, therefore, seems irrelevant to a discussion of the chemical potential of oxygen in the different rocks.

	Av.
Igneous rocks	1.23
Peridotites	1.73
Anorthosites	1.06
Basalts	1.18
Gabbros	1.93
Granodiorites	1.62
Granites	1.00
Keratophyres	1.75

The theoretical treatment of "Ramberg's principle" is based on the free passage of oxygen upward. Barth states (1948*a*, p. 43) that the higher oxides of iron are unstable at greater depths, where "they dissociate, with formation of oxygen, which also migrates upward." However, elsewhere Barth says that the oxygen might be regarded as a glassy solvent, in which "Ramberg's equilibrium" is established by migration of cations.

Barth seems to think that all rocks containing equal quantities of oxygen will take up the same volume. This assumption is obviously a misconception. He states that the volume relations of ordinary rocks are almost wholly determined by the oxygen ions. The most important factor in the volume relation of the rocks is, however, the packing index of the minerals. H. W. Fairbairn (1943) has handled this in an excellent way.

We now come to the subject treated in Barth's second paper, namely, the new method for petrographic calculations. In short, the method is based on the following: All ordinary rocks have a volume ratio of oxygen/cations = 92/8. Barth says, therefore, that, if one wants to compare rocks of equal volume, one can compare rocks containing the same amount of oxygen ions. As a standard comparison unit it is expedient to choose a volume comprising 160 oxygens, for in an ordinary rock nearly 100 cations are associated with 160 oxygens. Barth calls a rock unit containing exactly 160 oxygen ions the "standard cell."

It is evident that different minerals containing the same amount of oxygen, e.g., 160 oxygen atoms, may have very different volumes. This is valid even for minerals in the same metamorphic facies; comparing two minerals in the sanidinite facies—corundum, for example, with a "standard cell volume" of 160 oxygen atoms = 2,300 Å^3, cristobalite with a "standard cell volume" of 160 oxygen atoms = 3,520 Å^3. Barth says (p. 56) that some petrologists introduce entirely arbitrary assumptions, such as iron or alumina being constant during metasomatism. He himself assumes that the oxygen content is constant during the metasomatism. This assumption is, of course, also arbitrary. Barth thinks in this way to attain a method automatically satisfying the volume relations. This must be a mistake. We may, for example, consider the metasomatic transformation of dolomite into diopside. According to Barth's view, this process takes place by diffusion of Si^{4+} ions into the carbonate rock, followed by a replacement of the carbon in the carbonate and a diffusion of C^{4+} out of the rock. This leads to volume contraction of about 5 per cent.

In other cases, such as the transformation of spinel into olivine by replacement, the effective volume expansion would be nearly 20 per cent if the amount of oxygen is kept constant.

Barth, several times, writes carbon ions and silicon ions. It is, however, doubtful whether the four valent positive ions of carbon and silicon have any probability of existence in any terrestrial system. The C^{4+} ions have never yet been found, except at solar conditions. The ionization energies are extremely high in a system containing oxygen.

Barth's theories dealing with the inorganic cycle of oxygen do not seem to be in harmony with the rest of his paper. Here he pleads for the stationary oxygen "solvent." It is difficult to understand why the oxygen should be stationary because of its great ion radius (= 1.32 Å) and why potassium (ion radius = 1.40 Å) should be very mobile.

In any case, the diffusion coefficients in rocks are too low to cause any mass transport in the solid state over a great distance. The whole question is therefore a purely academic one.

However, I am not of the opinion that diffusion in the solid state is unimportant in metasomatic processes. In another paper I have compared metasomatism with the metabolism of living creatures, e.g., the oxygen flows with blood through arteries and capillaries, it diffuses through the cell walls into the cells. The excreta diffuses from the cells into the capillaries and flows to the kidneys and lungs.

Twenty-five years ago, Manjero Watanabé, in an excellent discussion of the diffusion process in geology, was aware of the impossibility of a long-distance diffusion in the earth. He writes (p. 134):

> Diffusion can perform metasomatism through fine passages even where circulation is difficult, while circulation furnishes the new material and removes the old substance and thus maintains the concentration necessary for diffusion. Without circulation, diffusion would be unable to accomplish much, because of its small velocity and of the exhaustion of material in the nearby source. Without diffusion, circulation would be unable to replace, the mineral, particle by particle; it can only dissolve the material on the surface of relatively large fissures and there deposits a new substance, but not beyond this zone.

In closing, I would mention that sulfide minerals commonly seem to have replaced oxide minerals without volume change. In these cases it is obvious that the oxygen has moved.

I am aware of the fact that this paper has been an entirely negative one, that I have not suggested any better method of solving the problems dealt with by Barth. Elsewhere, I have in some manner dealt with similar problems, and, until a better solution can be found, I prefer the more conservative interpretations, although I am aware of their possible shortcomings.

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THE DISTRIBUTION OF OXYGEN IN THE LITHOSPHERE: A REPLY[1]

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In the preceding discussion Mr. I. Th. Rosenqvist is not so much concerned with the distribution of oxygen in the lithosphere as he is opposed to "Ramberg's principle." This principle was demonstrated some sixty years ago by Willard Gibbs; now Rosenqvist is opposed to its being applied to the stability relations in the crust of the earth as was recently done by Mr. Ramberg.

In an attempt fully to understand Mr. Rosenqvist's standpoint I have carefully read his previous discussions (1947 and 1948). As far as I can make out, Rosenqvist's final, general complaint can now be reduced to what is contained in the following passages: ". . . the chemical activity of a crystalline substance is dependent not only on the temperature and pressure but also on the surrounding phases . . . traces of alien ions in a crystal lattice have considerable influence on the activity. . . . Nothing can be stated, therefore, as to the variation of the chemical potential of a mineral compound with pressure. . . . Ramberg's principle is valid only in a pure system in a vacuum."

Let us see how thermodynamics gives us the relation between chemical activity and pressure in a rock. To this end let us consider a volume unit of a rock over a range of temperature and pressure. The component phases of the rock are the various minerals and, at higher temperature, a "pore" liquid (or gas); these phases may contain traces of any of the existing chemical elements. At a given P-T condition we assume the phases to be at equilibrium, and the chemical activity of, say, oxygen is therefore the same in all phases. Suppose the rock contains quartz, as many rocks do. With increasing pressure the chemical activity, a, of oxygen in quartz will increase from a_0 to a_P according to the equation

$$a_P = a_0 e^{\Delta PV/RT} , \qquad (1)$$

where ΔP is the increase in pressure and V is the fictive volume of oxygen in quartz.

In all other component phases of the rock the activity of oxygen will increase according to the same formula (1). Since the fictive molal volume of oxygen may be somewhat different in the different phases, the activity of oxygen will not be exactly the same for all phases at the new pressure, P. Reactions will therefore take place in such a way that the small oxygen activity differences are eliminated and the new stable oxygen activity will be of a value intermediate between those given by formula (1) when applied to the components with the greatest and smallest V.

It seems to me that Rosenqvist, instead of considering the effect of the pressure, has discussed the effect of the addition of a reactive solvent to the rock powder. Now the "pore solution" may be regarded as a solvent, but it is hardly very reactive; and the claim that it may make "the reactions develop just opposite to the direction to be expected" will be received, I believe, with great skepticism by most petrologists.

Incidentally, similar thermodynamic equations have subsequently been applied to petrogenetic processes by Brown and Patterson (1948), who won the prize of the American Association for the Advancement of Science for their paper and whom Rosenqvist quotes with approval because they state that definite proof must await refined thermodynamical calcula-

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[1] Manuscript received March 3, 1949.

tions. But if Rosenqvist's assertion is true, that "nothing can be stated as to the variation of the chemical potential of a mineral compound with pressure," then it seems to me that the refined thermodynamical calculations are of no use.

To be sure, Rosenqvist's objections are not new to petrologists. In his latest textbook Niggli states that, although theoretical chemistry is based on ideal conditions which are never attained in nature, it would be quite wrong to conclude that it has no application. Niggli (1948, p. 375) says:

> Da der Mensch, eingespannt in das von ihm geschaffene Begriffsystem, gerne zur Einseitigkeit neigt, hat man under Bezugnahme auf die erwähnten Erscheinungen auch von einer Nictanwendbarkeit der Phasenlehre auf die Gesteinskunde gesprochen.
>
> Die Theorie geht von idealisierten Verhältnissen aus, die gar nicht erreichbar sind. Aber es wäre naturgemäss unrichtig, aus diesen Widersprüchen zu schliessen, die klassische Phasenlehre sei überhaupt nicht anwendbar oder es sei, ausgehend von den gleichen Prinzipien, unmöglich, sie anderen Voraussetzungen anzupassen. Hat er das Geschehen in grossgeologischen Räumen zu überblicken, so muss er zu den thermodynamischen Grundprinzipien zurückkehren.

Throughout his discussion Rosenqvist is concerned with questions of no pertinence to the article which he professes to criticize.

For example, he brings up the volume relations in the equation:

$$4Fe_2O_3 + Fe^{++} = 3Fe_3O_4 - 2e\ .$$

But this equation should not be discussed without considering the relations of FeO, and in my article I made special mention of the fact that "the possible effect of the 'reversed' specific gravity relations of hematite and magnetite will not be discussed here." Now an example involving the same problem has been treated by Ramberg (gypsum ⇄ anhydrite + water), and Rosenqvist may be referred to this discussion (1949*b*).

Likewise, the quixotic fight put up by Rosenqvist to prove that "[diffusion through crystalline minerals] is without any geological significance" falls outside the scope of the article discussed by him. In my article I made no assumption as to the mechanism of the migration of the atoms, and Ramberg (1949*a*) says that he "considers it unlikely that large-scale diffusion through solid rocks preferably takes place through the silicate lattices."

Rosenqvist misrepresents my view by assuming that "oxygen should be stationary because of its great ionic radius, =1.32 Å, [whereas] potassium, ionic radius = 1.40 Å, should be very mobile." My statement reads (p. 47): "Owing to the great bulk of the oxygen ions, they probably furnish a rather stationary medium." In this sentence I referred to the total bulk of the oxygen ions in the lithosphere (=92 per cent by volume), not to the ionic radius. I cannot remember having said that potassium is "very mobile," and I have certainly as yet never suggested any general relation between mobility and atomic radius.

The listing of the cations associated with 160 oxygens, as required by my scheme of petrographic calculations, does not imply that the cations necessarily have moved as such; nor does the listing of rock analyses in terms of oxides imply that the oxides exist as such. Ramberg says (1949*a*, p. 52):

> It can be shown that the mobility of the several particles in rocks increases with rising partial vapor tension of the particles. It seems very probable that large-scale migration through solid rocks takes place with the help of "gaseous" atoms and molecules adhered to the mineral surfaces or the mosaic fissures.

To me the picture of the mechanism of the diffusion is still uncertain and tentative. But the solution may perhaps now be found in the highly interesting structural control of the metasomatic processes as recently demonstrated by Newhouse (1949).

At this point it should be emphasized that thermodynamics tells nothing about the velocity and mechanism of a reaction. Thermodynamics gives us only the directions of the several reactions. The actual reactions may be slow or fast and may use different paths, but each reaction that takes place brings the system of rocks closer to equilibrium.

There are and will always be conservative geologists, but thermodynamics favors neither conservatives nor progressives; it simply presents us with the ideal equilibrium conditions; and observations in nature, not in theory, must decide whether or not equilibrium has been attained.

Rosenqvist favors what he calls a "conservative interpretation" and thinks that the upper part of the lithosphere is rich in quartz and feldspar for the same reason that the upper part of a milk bottle is richer in fat than the lower part. Now many serious students of the subject may look upon this as a very radical interpretation indeed. But in thermodynamics these words have no meaning. Thermodynamics simply states, as I have endeavored to show in my article, that equilibrium in the lithosphere is attained if oxygen accumulates near the top. And observations have shown that in the earth's crust oxygen consistently decreases with depth.

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THE DISTRIBUTION OF OXYGEN IN THE LITHOSPHERE COMMENTS ON BARTH'S REPLY[1]

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I may summarize my objections as follows:

1. I do not think that it is correct to apply the equations of Gibbs in a complex system (i.e., in a rock).

2. "As long as it is unknown how the solubility of different substances in different crystal lattices are related to pressure, nothing can be stated . . . as to the variation of the chemical potential of a mineral compound with pressure." Consequently, refined thermodynamical calculations and laboratory investigations are necessary before anything can be established as to the variation of the chemical potential with pressure.

3. When Barth prefers to term my discussion of the unimportance of diffusion "quixotic," it must be remembered that diffusion (through the crystals or through the intergranular film) is the *only* path of transport where an investigation of the variations of chemical potential is of any interest.

THE USE OF OXYGEN AS A REFERENCE SUBSTANCE IN PETROGRAPHIC CALCULATIONS: A SUPPLEMENT AND A REPLY[2]

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In recent years much attention has been given to the problem of granitization and, generally, to metasomatic material transfer in rocks and reactions in the solid state. Elsewhere I have reviewed some of the recent contributions to this subject (Barth, 1948*b*). But nobody, I believe, has given credit to the first French papers that actually initiated this school of thought fifteen years ago.

Perrin (1934), extrapolating from metallurgical observations, formulated the hypothesis that metamorphism (metasomatism) was due to diffusion in the solid state in opposition to the existing theories that called for intervention of liquids or gases. Likewise, he suggested (1935) that granite was the product of such diffusions and had never crystallized from a liquid magma. In regard to material exchange in solid rocks

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[1] Manuscript received March 14, 1949.

[2] Manuscript received March 3, 1949.

Perrin writes (personal communication 1948, translated from French):

> The first question is to find out if the atoms of certain chemical species remain in place. Lapadu-Hargues [1945] in his study of metamorphism first thought that silicon and aluminium did so. We [Perrin and Roubault, 1946] pointed out to him that his own statistical methods demonstrated an appreciable variation in the ratio Si/Al and that nothing indicated that any of these species remained stationary. In the meantime you [Barth, 1948*a*] reached the conclusion that the solvent ions to a large extent were stationary and that the cations moved through this solvent. A priori this seemed a probable hypothesis to me in view of the enormous preponderance of oxygen in the lithosphere. . . . Subsequently, Lapadu-Hargues [1948] came to the same conclusion. On a small scale, however, some oxygen diffusion may occur, e.g, in the case of pyrite replacing oxides, etc. . . . It seems to me that all one may say is that, because of its very great concentration, oxygen is without doubt the atom that diffuses the least and plays the role of a solvent. But it does not mean that the frame of oxygen remains unchangeable, the shape of the frame certainly changes. Perhaps you have generalized too much in saying that the volume remains constant during metasomatic exchange. No doubt there are cases where observations would seem to demonstrate that the changes in volume have been zero or insignificant. But Brajnikov [1945] has pointed to the variation in the number of oxygen atoms per cubic centimeter of the different minerals. The result is that if, e.g., a crystal of orthoclase during granitization devours a crystal of hornblende or quartz, then the oxygen lattice will expand; but if quartz replaces a plagioclase the lattice will contract. . . . What takes place on a small scale may take place on a large scale. Brajnikov directs attention to the fact that, although granular rocks show approximately the same value for the total concentration of oxygen per cubic centimeter, the corresponding value is larger for detritic rocks. . . . It seems to me as an automatic conclusion, if no notable diffusion of oxygen has occurred, that metasomatism and granitization are generally accompanied by a total dilatation of the oxygen lattice and by an increase of the volume of the rock. This is a verification of the hypothesis of metamorphism as a generator of folding as I announced in 1935.

Some of the above conclusions of Mr. Perrin will be discussed presently. First, I want to quote some statements by Rosenqvist in a preceding article in this *Journal* (p. 422): "This assumption [that the oxygen content is stationary during the metasomatism] is, of course, also arbitrary. Barth thinks in this way to attain a method automatically satisfying the volume relations. This must be a mistake."

In this connection reference should be made to Brajnikov's article (1945), in which he has calculated the absolute amount of oxygen in the various minerals and rocks expressed in terms of the number of oxygen ions contained in a cubic centimeter of the substance. I have shown (Barth, 1948*c*) that Brajnikov's article supports my point of view. I shall repeat the salient points here.

As can be seen from Brajnikov's tables, gabbro and granite have very nearly the same number of oxygen ions per cubic centimeter. Many of the ordinary rock types are transitional between gabbro and granite and therefore also contain the same number of oxygen. Granitization of gabbroic and, generally speaking, of metamorphic rocks takes place, therefore, without appreciable change in volume if the oxygen content is kept constant. This is a fact and no mistake.

It should be remembered that in previous papers I have repeatedly emphasized that only in the ordinary silicate types of rock belonging to the same mineral facies are the volume relations, in this manner and in their first approximation, determined by the oxygen ions. It is therefore completely irrelevant when Rosenqvist discusses the volume relations of corundum and cristobalite.

Perrin's discussion raises a different problem. He thinks that complete granitization of sedimentary rocks involves a volume expansion of about 8 per cent, and he remarks that this corresponds to a linear expansion of only 2 per cent, a rather small amount, which explains the *mise en place* of many granites showing no derangement of the adjacent sedimentary strata. On the other hand, if one looks at large-scale granitizations comprising great volumes of the crust of the earth, like the roots of mountain chains, then the effect of expansion becomes of importance—indeed, Perrin thinks that it may be a factor in orogenesis.

I agree with Perrin that, particularly if transitions from one mineral facies into another takes place, for example, from the facies of sedimentary rocks into that of a granite, then the volume differences become of importance. (But an increase of 8 per cent is in my opinion not well supported.) It may be recalled that, for example, the transition from eclogite facies into amphibolite facies includes the important reaction

$$\text{jadeite} + \text{omphasite} \rightleftharpoons \text{plagioclase} + \text{hornblende},$$

which is accompanied by a volume increase of 15–20 per cent.

Finally, I should like to clarify the meaning of a "stationary" oxygen medium. Let us specifically consider the granitization process: (1) One cubic centimeter of the educt has approximately the same number of oxygen ions as 1 cubic centimeter of the product. (2) The metasomatic process has taken place volume by volume. (The volume law may be questioned, but from many places of the world it has been shown to hold, and from nowhere do we find decisive evidence to the contrary.) Consequently, we need not *in our calculations* consider any migration of oxygen; therefore, we may say that oxygen is stationary. But if, for a brief moment, we look into the actual mechanism, the picture becomes different. Thermodynamic equilibrium at a phase boundary does not mean a state of no motion, it means that an equal number of atoms or molecules migrate both ways, the net result of the transport being zero. Oxygen ions may move about in the rocks, but, following no preferred direction, they move statistically, usually causing no noticeable concentration in any part of the rock. They may act as carriers for silicon or aluminium, however; for the strongly bound silicon ion, incarcerated, as it were, in the center of an oxygen tetrahedron, may not easily break loose without the accompaniment of one or more oxygens. But the net result is that silicon moves, while the oxygen content remains "stationary."

Therefore, I do believe that the use of oxygen as a reference substance in petrographic calculations is justified and may prove fruitful in the study of, e.g., granitization, basic fronts, and similar phenomena.

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THE PRESERVATION OF ANTARCTIC ICE SPECIMENS: A DISCUSSION[1]

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The purpose of A. D. Howard's paper, "The Preservation of Antarctic Ice Specimens," which appeared in the January, 1948, issue of the *Journal of Geology*, is to demonstrate the feasibility of large spatial and temporal separation of glaciological field work and laboratory investigation of collected ice specimens.

Howard's thesis is largely fallacious because of the fact that the structure of ice aggregates is very sensitive to environmental change, particularly to variation of temperature and stress, and probably also to hydrostatic pressure. I have observed great changes of grain size due to recrystallization in thin sections of ice stored for no more than 1 or 2 weeks at −5° to −10° C. This ice was fine grained (2–3 mm.) and had been subjected to stress prior to sectioning. Recrystallization is slower, the larger the grain size, the smaller the stress prior to sectioning, and the lower the storage temperature.

Thus the structure of ice specimens taken from stressed portions of a glacier, from drill

[1] Manuscript received January 31, 1949.

cores or crevasses, should be examined soon after collection or stored at very low temperatures, for example, on dry ice (CO_2). Strong cooling, however, tends to crack larger specimens, rendering them useless for ice-mechanics work.

Wet ice can be frozen for sectioning, but it is questionable whether it may be frozen for subsequent mechanical testing. It remains to be proved that freezing and rethawing does not significantly upset a phase equilibrium which has been established during many years at pressure melting temperature.

The high sensitivity of ice structures to external conditions promises to be a valuable source of information on their near and remote past history. Addition of the impress of storage tends to obliterate the picture of at least the near past physical history. If laboratory work is to be done on natural ice, laboratory facilities should be provided very near to the source of the ice to be examined.

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THE PRESERVATION OF ANTARCTIC ICE SPECIMENS: A REPLY[1]

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The best procedure for studies of natural ice is on-the-spot investigation by specialists with all necessary equipment and with adequate time at their disposal. Unfortunately, this is impossible on Antarctic expeditions unless the studies are conducted in the immediate vicinity of base camp. Travel farther afield generally involves arduous sledge journeys, in which the problem of weight necessarily restricts the amount of equipment that can be taken along. The airplane and tracked vehicle may eventually eliminate this restriction. Unless the ideal combination can be arranged, however, and until we understand better the situations in which changes occur and the magnitude of the changes involved when ice specimens are removed from their natural habitat, it seems advisable to encourage wide sampling, even if the detailed examinations must be made later.

Dr. Bader opposes any delay in examination of ice specimens because "the structure of ice aggregates is very sensitive to environmental change. . . ." His entire criticism is based on the observation that "great changes of grain size due to recrystallization" took place within 1 or 2 weeks in thin sections of previously stressed ice which had been stored at −5° to −10° C. One might conclude, therefore, that ice removed from a stress environment changes so rapidly, regardless of temperature control, that immediate examination is mandatory. Yet Bader says that when specimens from stressed portions of a glacier cannot be examined soon after collection, they should be "stored at very low temperature." Thus Bader himself seems to concede that delay is feasible, if proper temperature precautions are taken. In view of this concession, I find it difficult to understand his strong condemnation of my proposal.

Recrystallization involves molecular transfer. Transfer in both liquid and vapor phase falls off with the temperature. For example, the vapor pressure of ice at −30° C. is less than seven one-hundredths of that at 0° C., and at −40° C., only a little over two one-hundredths. Storage at low temperatures, therefore, should retard very appreciably any changes due to recrystallization. Nor is it likely that there are significant changes other than those due to recrystallization. Ice under stress, as in a glacier, does not rebound to its original form when the stress is removed, nor do grains deformed by crystal gliding spring back to their original shape. The stress characteristics remain in the sample and may be modified only by recrystallization. But if storage at very low temperature can vastly retard recrystallization, then no unnecessary haste in examination of properly stored specimens is required. The effectiveness of low temperature in retarding increase in grain

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[1] Manuscript received February 14, 1949.

size is suggested by the observations of Wright and Priestly (1922, p. 124), who found "no tendency toward an increase in mean size of glacier grain at high pressures (great depth) in Antarctic glaciers."

One gathers from Bader's observations of change in grain size in sections stored at −5° to −10° C. that the temperatures were not low enough or that additional precautions against recrystallization were needed. Seligman (1948, p. 191) cites the interesting experiments of De Quervain, in which delicate snow crystals, immersed in paraffin oil at a temperature of −10° C. showed no alteration after periods of months. Alteration did take place, however, at −5° C. Demorest (1940, p. 127) stored specimens for months in kerosene at low temperatures but unfortunately did not publish on the results.

The specimens which I brought back from the Antarctic included not only samples of "stressed" ice, but also of névé and sublimation crystals. With regard to specimens other than glacial ice, a few observations by Wright and Priestley seem pertinent. They found (1922, pp. 79, 108) that certain megascopic structures in ice actually became more pronounced—therefore more easily studied—when the samples were subjected to ablation for periods up to 2 weeks in temperatures slightly below freezing. They found, too (p. 103), that, in contrast with the situation in temperate climates, changes in surface névé in Antarctica required a considerable lapse of time. They report that, after 2 years, at a depth of about 2 feet in the Ross Barrier, the individual grains still preserved the original subangular polygonal form. Such observations would seem to suggest that, if specimens of névé were stored under proper temperature conditions, the original characteristics might remain unaltered for considerable lengths of time.

None of the above comments are to be construed as arguments against research on possible changes in natural ice when it is removed from its habitat. There is also need, however, for additional research on improved storage techniques to insure a minimum of change in ice specimens which cannot be examined immediately.

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ALBITE–WATER SYSTEM IN THE GRAVITATIONAL FIELD: A COMMENT[1]

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The effect of the gravitational field on chemical equilibrium is usually negligible in laboratories, and the subject, therefore, is little discussed in the physicochemical literature. In petrology and geochemistry, however, any effect of the gravitational field—one of the principal properties of the earth—should be considered. With that in mind, the writer has emphasized the influence of gravity upon large-scale chemical equilibrium in the upper shells of the earth (1946–1949); later Barth dealt with the same phenomenon (1948), as did Verhoogen in his interesting paper on the "Thermodynamics of a Magmatic Gas Phase" (1949*b*) and in a discussion (1949*a*).

Verhoogen derives the thermodynamically correct relation between composition of a solution and height in the gravitational field (eq. V13, p. 129, 1949*b*) and applies this equation to the hydrous albite melt investigated by Goranson (1938).

At 1,000 bars and 900° C. an albite melt will dissolve 6 per cent water (provided that the partial water pressure corresponds exactly to the total pressure). From the data available and equation V13, Verhoogen finds that at 2,900 meters higher in the column of solution the concentration of water will be 8.2 per cent, because

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[1] Manuscript received February 28, 1949.

of gravitative differentiation. At this higher level the hydrostatic pressure in the solution is 500 bars, and Goranson's experiments show that only 4 per cent of water can dissolve in the melt. Verhoogen therefore arrives at the following conclusions (1949*b*, p. 130):

> The effect of the gravitational field is thus to increase very noticeably the saturation concentration. . . . In the same manner, a saturated solution at 900° C. and 500 bars (4 per cent) would be in equilibrium with a solution located at 2,900 m below and containing only 2.8 per cent of water instead of 6 per cent. . . . It follows that if the effect of the gravitational field were to disappear suddenly, as if, for instance, the upper and lower halves of the magma column were brought mechanically to a common mean level, the upper half would be strongly supersaturated and the lower half strongly undersaturated with respect to the new position.

These conclusions, if correct, would indeed be of the greatest geological importance; for it is only necessary to provide a sufficiently thick layer—or high column—of silicate melt–water solution in the earth and the upper part would easily dissolve 20, 30, or 50 per cent of water without any supersaturation. As such water-rich silicate melts must have a low solidification temperature, we could easily explain the formation of many granites and pegmatites which now are very problematic.

Imagine, now, that we follow the principle of Verhoogen and choose an albite melt with 6 per cent of water at 900° C. and 1,000 bars. According to Goranson, this system will be a homogeneous melt. Going deeper and deeper down in a column of such a solution, the concentration of water should decrease according to equation V13 (Verhoogen, 1949*b*), and at a certain depth the solution will contain, say, only 1 per cent of water. It is well known from Goranson's experiments that such a solution is strongly unstable and will crystallize to albite under expulsion of water (the same is true for the melt with 2.8 per cent of water). According to Verhoogen's conclusions, however, this rather "dry" albite melt will be a stable liquid at as low a temperature as 900° C. because it is situated at the bottom of the solution column. It may be tempting to believe that many deep-seated rocks are formed from such "dry" but low-melting magmas. Before we do so, however, we should study the reliability of the conclusions a little further.

To be sure, it is not necessary to imagine that the effect of the gravitational field disappears, or to split the column into halves in order that the 8.2 per cent of water in the upper part and the 2.8 per cent in the lower part shall represent strong supersaturation and undersaturation, respectively. The 8.2 per cent of water in albite melt at 900° C. and 500 bars makes a strongly supersaturated solution, independent of any continuation with a subjacent melt in the gravitational field. Likewise, an albite melt with 2.8 per cent of water at 900° C. and 1,000 bars is strongly "undersaturated" with respect to water, entirely independent of a superincumbent column of melt.

The "supersaturated" melt will split up into melt and water-rich vapor. The "undersaturated" melt will form albite crystals and a new melt with 6 per cent of water.

When considering the effect of gravity upon such systems, one should remember that there are two equally important requirements to be fulfilled, namely, (1) equation V13 must be followed in homogeneous phases in order to insure chemical equilibrium from top to bottom in the column, and (2) the concentration of water in the albite melt cannot surpass certain limits at any level in the column, otherwise water vapor will develop or albite crystals form.

Apparently Verhoogen does not consider the second requirement; and therefore he concludes that the gravitational field alone would be able to decrease or to increase the saturation concentration as postulated.

It is easy to show that the albite melt–water column in the gravitational field at 900° C. will behave as follows: the upper part of it will expel water-rich vapor, which will gather above the top of the melt; and the lower part will yield albite crystals, gathering in a bottom layer. When complete equilibrium is attained, the upper part of the melt will just be saturated with water under the pressure there—say, 1,000 bars and 6 per cent water (if 500 bars and 4 per cent water is used, no melt will exist). The concentration of water down in the melt will follow equation V13, so that concentration of water decreases with depth. Now, since a melt with 6 per cent of water at a temperature of 900° C. is near the crystallization point (Goranson, 1938, fig. 4, p. 85), the melt column will be very short. At the bottom of that short column the concentration of water may be, say, 5.9 per cent or just on the crystallization point. Below that level neither melt nor water vapor exists except perhaps as minute pore fillings in the albite crystalline aggregate, which is stable farther downward.

In conclusion, the gravitational field is unable to stabilize systems which are otherwise chemically unstable at the pressures and temperatures in question. The effect of the gravitational field is to change hydrostatic pressure and to change bulk chemical composition at different levels in the earth's crust.

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REVIEWS

Glaciological Research on the North Atlantic Coasts. By HANS W:SON AHLMANN. ("Royal Geog. Soc., Research Ser.," No. 1.) London, 1948. Pp. 83; figs. 44.

This publication of 83 closely printed pages, 44 figures, and numerous tables—the first volume of a new research series issued by the Royal Geographical Society—provides the most succinct and best-organized summary of H. W. Ahlmann's extensive glaciological studies yet to appear in English. Dr. Ahlmann, professor of geography at Stockholm University, has established himself as one of the world's best authorities on glacier nourishment and wastage through the researches herein summarized. His work, well known in the United States and Canada through earlier extensive writings and a 1947 lecture tour in these countries, will gain further well-deserved recognition through this publication.

Studies covering a period of twenty-eight years, 1918–1946, and dealing with six separate areas bordering the northernmost Atlantic Ocean are reported. The areas and glaciers investigated are the Styggedal and associated glaciers in the Jotunheim district of southern Norway; the Kårsa Glacier of Swedish Lapland; the ice caps of Northeast Land, the northeastern island of the Svalbard Archipelago; the ice of Isachsen's Plateau and the Fourteenth of July Glacier in West Spitsbergen; the Fröya Glacier on Clavering Island off the east coast of Greenland; and two outlet glaciers of the Vatnajökull—the Hoffell and Heinaberg—in Iceland.

This work has profited from the assistance and collaboration of a host of able colleagues best known among whom are H. U. Sverdrup and Sigurdur Thorarinsson. Ahlmann's wide acquaintance with and generous use of a great body of foreign glaciological literature, not well known to most American glaciologists, is a further asset. The subject matter is presented under the following headings.

Accumulation.—Although snow is the major source of nourishment, hail, hoarfrost, rime, and local rains that freeze before draining away are also contributors. Wind drifting of snow is favored by light snow, flat topography, low temperature, and strong wind. However, Ahlmann warns that it is a great error to think that all glaciers must have leeward orientation. Jotunheim is cited as a district which receives considerable windward nourishment in the form of hoarfrost, and on Vatnajökull the largest development of glaciers is to the windward. Recognition of the role played by freezing of rain and meltwater in nourishment, firnification, and icification will be approved by most glaciologists. Of the 485 mm. of water percolating into the snow on Isachsen's Plateau in 1934, 145 mm. changed to ice and iced firn, while 340 mm. sank through the firn and drained away. In a cold polar glacier, nearly 100 per cent of the water percolating into the firn must freeze, but its absolute volume is smaller than on temperate glaciers.

The accumulation area is that part of a glacier in which accumulation surpasses ablation. On the Fourteenth of July Glacier, 27 per cent of the total accumulation fell in the accumulation area and 73 per cent in the ablation area. By contrast, Hoffell Glacier received up to 89 per cent of its total accumulation in the accumulation area. This serves to emphasize a point repeatedly made by Ahlmann that the morphology of a glacier is of great significance in determining its regimen and behavior. One would rightly conclude from the above figures that Hoffell Glacier is the much more energetic of the two. Curves showing the amount of accumulation at various altitudes on different glaciers also display marked individuality. Accumulation usually increases with increasing altitude up to a point where it slackens off, and the rate of increase and the slack-off point for various glaciers are markedly different, depending upon climatological environment and orographic situation.

Many details of interest to students of physical glaciology, such as thermal conditions and structural features within the firn fields studied, are also summarized in this section of the article.

Ablation.—Ablation is defined as the glacier-wasting process and, in addition to melting and evaporation, is said to include calving and corrasion by the wind. The last two processes have not usually been considered part of ablation by American glaciologists, and this definition may raise a few eyebrows. Ablation is chiefly a surface phenomenon, but it may also be internal, for wasting occurs on the walls of crevasses and in englacial or subglacial hollows and tunnels. Since meltwater may be entrapped in the glacier and refrozen, it is necessary to distinguish between gross or total ablation and net ablation, that is, the quantity of water wholly removed from the glacier by drainage, evaporation, corrasion, and calving.

The treatment of various factors and conditions controlling ablation appears to permit the following conclusions:

1. Radiation, temperature conduction, condensation, and evaporation are all factors in ablation, but radiation and temperature conduction are of primary importance, and evaporation is relatively insignificant in most environments. Condensation can be of considerable effect under proper conditions. The relative efficacy of radiation and temperature conduction depends upon many variables, including latitude, altitude, season, and climatological environment.

2. In general, the part played by radiation in causing ablation increases with latitude and altitude. Even on cloudy days in the arctic the effective radiation may be higher than on clear days. This is not true of lower latitudes and apparently comes about because of reduced albedo and the thinness of the cloud cover in far northern areas.

3. The relative influence of radiation on ablation decreases as the season progresses, and the effect of temperature increases. Late in the season when the snow line reaches its highest position and becomes the firn line, ablation is said to be largely or wholly determined by temperature conduction.

4. With decreasing altitude, temperature change and wind effects are increasingly effective in causing ablation. Detailed studies on Hoffell Glacier show that an increase in temperature has a greater absolute effect in low than in high altitude. An increase in air circulation nearly always accelerates ablation.

5. Ablation on ice is larger than on snow and by a greater degree, depending upon the importance of radiation in causing ablation. This comes about because of the greater albedo of snow.

6. The decrease of ablation with altitude is much greater on ice than on snow, 340 mm. per 100 meters as compared to 70 mm. per 100 meters in the instance of the Fourteenth of July Glacier.

7. With increasing elevation, summertime ablation accounts for an increasingly larger part of the total ablation. On Hoffell Glacier, June to August ablation accounts for 55 per cent of the total at 72 meters altitude, 60 per cent at 400 meters, and 78 per cent at 1,100 meters.

8. Gross ablation values may differ greatly from year to year. On Hoffell Glacier the 1937 ablation was only 59 per cent of that for 1936.

9. The morphology of a glacier, that is, its shape, slope, and area distribution, has a great effect on both accumulation and ablation. For example, a gently sloping ice surface permits notable shifts of the snow line with only modest meteorological change, and such glaciers are necessarily most sensitive to climatic change.

10. Meteorological conditions in the area of the snow line strongly influence the glacier regimen.

Firn and glaciation limits.—The discussion of firn limit (or line), snow line, and glaciation limit presented in this section will be of particular interest to persons not already familiar with these concepts. Glaciation limit is not defined but apparently refers to the level at which glaciers form in any particular region. It is emphasized in this treatment that the summer isotherm at the glaciation limit gives a measure, or at least an indication, of accumulation at that altitude; the higher the temperature, the greater the accumulation.

Regime.—The total accumulation volume of a glacier during one accumulation season and its gross ablation volume during the following ablation season constitute the regime, or material balance, of a glacier. If accumulation exceeds ablation, the regime is positive. The budget year for a glacier starts with the autumn accumulation, usually in September or October, and the sums on the balance sheet are expressed in cubic meters per square kilometer. The sum of total accumulation and total ablation is the glacier's balance-sheet total. Regime diagrams, consisting of a net ablation curve and an accumulatian curve for areas at specified altitude intervals, show at a glance the glacier's state of health. Climatological conditions in May and September exert particular influence on a glacier's regime; for lengthening or shortening of the ablation season has more effect on total ablation than has any other change. Temperature variation appears to be the basic and decisive factor controlling changes in ablation. Furthermore, changes in ablation have greater influence on glacier regime than do changes in accumulation; at least, this is true on the Fourteenth of July and Hoffell glaciers.

Assembled data on a number of glacier regimes show that accumulation plus ablation in cubic centimeters of water per unit area at the firn limit is roughly the same as the average for the whole glacier. Thus a glacier's balance-sheet total equals the accumulation plus the ablation at the firn limit multiplied by the area of the glacier. If this relation holds in general, then observations to establish the total regime can be made at the firn limit. Meteorological observations at the firn limit also provide data on the climatic elements most effective in influencing a glacier's state of health.

Rate of movement.—Lines joining points of equal average rate of movement on a glacier's surface are "isotachytes," and isotachyte maps of several glaciers are presented. Admittedly, some of these are crude approximations, that of the Fourteenth of July Glacier, for instance, being based on only seventeen known points. Even so, the maps present principles of ice movement in striking fashion.

On Styggedal Glacier the rate of movement in the ablation area is 3–21 per cent greater in summer than in winter, and the difference is least when the rate is greatest. Great variations in movement of the Hoffell Glacier in different years and from place to place lead to the suggestion that waves of different velocity of movement proceed from the firn area to the glacier tongue. The rate of movement is to a large extent an expression of the amount of the glacier's regime or, more properly, the size of its balance-sheet total. If the latter is large, movement is relatively fast, even though the glacier may be receding, for much ice must be transferred from the accumulation area to the ablation area.

Classification of glaciers.—Classifying glaciers has long been an unsatisfactory business, and Ahlmann faces squarely up to the problem by advocating three separate classifications, namely, morphological (external features), dynamic (degree of activity), and geophysical (temperature and physical properties of the firn and ice). His morphological classification includes the major elements of earlier classifications and recognizes essentially three major divisions; ice sheets and caps, confined glaciers, and piedmont glaciers and related types. However, this arrangement is put on a more precise basis than in previous morphological classifications by the construction of percentile area distribution diagrams or curves for each of the principal types. These diagrams appear to be reasonably distinctive and should be a considerable aid in placing a glacier in its proper pigeonhole.

On a dynamic basis, glaciers are classified as active, passive, or dead. The activity or passivity of a glacier depends upon its depth, speed of flow, and balance sheet. The rate of movement of an active glacier is large in relation to its size and slope and that of a passive glacier is small in the same relations. Even a receding glacier may be classified as active if it has a large amount of ice to move from the accumulation area to the ablation area and is flowing at a rapid rate to accomplish this task. Glaciers on the south side of Vatnajökull have long had a negative balance and are consequently receding, but they lie in an area of intense glacial energy and are properly classed as active.

The definition of dynamically dead ice is as follows. "When the negative regime has continued until the movement of the remaining ice is no longer caused by the transfer of material from an accumulation area, but is determined only by the slope of the bed, the ice is said to be dead." This statement might be modified or extended to include Demorest's concept of pressure-controlled ice flowage, which, along with gravity flow, was used by Matthes as a basis for a different dynamic classification of glaciers. Ahlmann also observes that ice is dead from a climatological viewpoint when it no longer receives nourishment from an accumulation excess above a firn line but exists on its bulk alone. In a large glacier, several thousand years may elapse between climatic and dynamic deaths.

The geophysical classification is well known from Ahlmann's earlier writings and distinguishes temperate glaciers and polar glaciers with a subdivision into high-polar and subpolar in the last category. These distinctions are made largely on temperature regimen and physical condition of the firn and ice.

Variations of glaciers.—The glaciers investigated have been receding and shrinking because of a long-continued climatic amelioration, considerably intensified in the past decade. The present rapid, partly catastrophic, shrinkage stands out as the last stage in a decline which set in following the maximum advance of these glaciers within historical time. This advance may also have been the maximum within postglacial time. Throughout much of the North Atlantic region corresponding double-end moraines mark this maximum ice extent. This position has been reached in different areas at one or the other of the several "Hochstand" periods widely recognized as occurring (*a*) early in the seventeenth century, (*b*) about the middle of the eighteenth century, (*c*) early in the nineteenth century, and (*d*) about the middle of the nineteenth century. In Spitsbergen, at least, it is also recognized that the glaciers shrank to a fraction of their present size during the postglacial warm, dry period and a corresponding shrinkage is believed to have occurred in other areas. Similar behavior has long been recognized in Alaska. American glaciologists will be interested in a reference to work (unpublished?) by A. Hamberg in which he suggested as early as 1896 that the present glaciers of the Kvikkjokk district, Sweden, had probably been regenerated after the postglacial warm, dry period. This, of course, parallels Matthes' "Little Ice Age" in the Sierra Nevada.

Thorarinsson concludes from extended analysis of the literature that marginal variations of glaciers in recent centuries have in all probability been more or less concurrent all over the globe and, further, that the "Hochstand" in the middle of the nineteenth century has been demonstrated in most glacier districts of the world. The significance of such a conclusion is obvious with respect to any theory of glaciation requiring alternate ice ages in the Northern and Southern hemispheres. Although present glacier deterioration appears to be world-wide, the rise of sea level is so slow that the great Greenland and Antarctic ice sheets cannot be melting as rapidly as most of the smaller glaciers are, according to Ahlmann.

The present climatic amelioration in the North Atlantic features a rise in temperature of both sea and air. Meteorological records in Norway and Svalbard show that the temperature has increased incomparably more since the middle of the nineteenth century than at any other time in the last two hundred years. In general, the winter temperature has risen the most. As previously shown, the recession and shrinkage of glaciers around the North Atlantic is due mainly to increased ablation, which, in turn, is attributed chiefly to a rise in temperature. This temperature change has been most effective through lengthening of the ablation season in spring and autumn. The "recent climatic improvement" seems to have been brought about largely by an intensification of atmospheric circulation which produces an increased supply of warm air from the south.

The reader must guard against being lulled into a false sense of accuracy and precision by Ahlmann's many impressive diagrams and tables. It is clear that the author has no delusions on this score, but apparently he does feel that the magnitude of the sums involved reduces the errors to insignificance. Occasionally a more specific demonstration that this is true would be desirable and may possibly be supplied by reference to more complete treatments by Ahlmann and his associates in the various volumes of the *Geografiska Annaler*. Literally hundreds, if not thousands, of pages in that journal are devoted to the matters treated here in abbreviated form. More specific reference to that extensive literature would help the interested student in obtaining further information on matters barely mentioned in this summary. For instance, one would like to know how the thickness of ice in a glacier is calculated from its speed of movement, amount of ablation, and distance from the margin (p. 57). Further elucidation of the glacial limit and isoglacihypse lines might be furnished by a reference to the pertinent literature. In addition, such statements as "The ablation is now (*in the fall*) largely or wholly determined by temperature conduction," which are not obvious in themselves, might be more readily acceptable if reference to a detailed treatment were given.

The above are minor points, obviously arising chiefly from the necessary brevity of treatment. This work is truly a fine contribution, and the English-reading geologist can be grateful to Dr. Ahlmann for his internationalism and ubiquity in the field of languages as well as for his sterling scientific performance.

R. P. S.

An Introduction to the Theory of Seismology. By K. E. Bullen. Cambridge: At the University Press; New York: Macmillan Co., 1947. Pp. xiii+263; figs. 43. $4.00.

The legitimate object of any technical writing is the transfer of information from one who knows more about a given subject to one who knows less. Accordingly, one of the first problems to which an author should direct his attention is to define the audience to whom he intends to address himself and then to formulate his presentation in such a manner that the majority of that audience is able to follow without excessive difficulty what is being said. Within the chosen framework and at the selected level of technicality, the treatise should be as nearly as possible logically self-sufficient and complete.

The present book is a treatise on the theory of seismology, and its author is professor of applied mathematics at the University of Melbourne and a former student and research associate of Professor Harold Jeffreys, of Cambridge University. The audience to whom the book is addressed is said to comprise "observatory seismologists" and "university undergraduates who are interested in seismology as a branch of applied mathematics." This, in effect, constitutes two separate audiences, whose needs, interests, and backgrounds are somewhat divergent; and the one which appears to have received the greatest attention is the latter, namely, the university students of applied mathematics.

The logical outline of the book is good. It begins with the theory of elasticity and proceeds to a discussion of vibrations and waves and thence to body and surface elastic waves. Then follow applications to seismology. This includes the theory of seismic rays in a spherically stratified earth model, and the amplitudes of the surface motion due to seismic waves. Following this is a brief development of the theory of the seismograph. Then the construction of travel-time curves is discussed, and deductions from seismic data are made concerning the layering and physical properties of the interior of the earth. Final chapters deal with the occurrence of earthquakes and miscellaneous seismic topics.

One of the severest tests of the book concerns the manner in which the various theoretical and observational considerations and data have been brought together to disclose the nature of the earth's interior. The handling of this problem has been particularly well done. The author has displayed great skill in deriving the maximum unambiguous information concerning the earth's interior and has exercised commendable restraint in not letting his deductions transgress his data.

The weakest aspect of the book, in this reviewer's opinion, is the somewhat inhomogeneous manner in which different parts are written and its failure to meet the needs of its potential readers. The chapters on elasticity and wave theory in particular are much too tersely written to be read without excessive labor and expenditure of time by any but accomplished students of applied mathematics. Since paper is one of the cheapest of commodities and human time one of the dearest, this waste of the reader's time, which could easily have been avoided by expanding the development into a more readily intelligible form, does not appear to be justified. This comment is even more pertinent with respect to repeated instances wherein important equations are stated without derivation and it is remarked that their derivation is left as an exercise for the reader.

Since the book is not properly self-sufficient, it would have been helpful if citations of significant original literature had been given. The citations given are limited mostly to a dozen or so general treatises, only one of which (*Handbuch der Geophysik*, Vol. IV) is in a language other than English. References (without literature citations) to the work of individuals occur throughout the book, but almost every alternate reference is to Jeffreys. From this one finds it difficult to escape the impression that nearly all the world's seismic researches have been published in English and that, of these, about half have been the work of Jeffreys.

To summarize: While this book is unquestionably a valuable addition to one's reference library on theoretical seismology, it falls somewhat short, for reasons given above, of adequately meeting the needs of the audience to whom it is addressed, and hence it fails to achieve the status of the outstanding text that it otherwise might have been. It is to be hoped that in future editions the book will be considerably expanded with these considerations in view.

M. KING HUBBERT

Bulletin of the Oceanographical Institute of Taiwan, nos. 1–3 (November, 1946; April, 1947; and June, 1947). In co-operation with the National Taiwan University and Institut de Biologie de Shanghai. Taipei, Taiwan, China: Oceanographical Institute of Taiwan.

Taiwan, also known as Formosa, is the home of a newly established institute of oceanography. Founded by the Chinese government following World War II, the institute has issued three numbers of the *Bulletin*. These, dated November, 1946; April, 1947; and June, 1947, contain the following papers:

No. 1.—(1) "Observations on the Coral Reefs of Penghu Islands, Taiwan," by T.Y.H. Ma, I. Hayasaka, and S. Kawaguchi; and (2) "On Fossil and Recent Brachiopods of Formosa," by I. Hayasaka.

No. 2.—"Submarine Valleys around the Southern Part of Taiwan and Their Geological Significance," by Ting Ying H. Ma; (2) "The Cause of Uplift of Eastern Asiatic Continent during the Middle Pleistocene Time," by Ting Ying H. Ma; (3) "Is There Any Particular Feature in the Marine Neogene Fauna of Taiwan?" by Ichirô Hayasaka; and (4) "Marine Molluscan Fauna from the Tashufang District, Kaohsiunghsien," by Sueo Kaneko.

No. 3.—(1) "The Upheaval Movement in Eastern Asia at the End of the Tertiary and the Cause Thereof," by Ting Ying H. Ma; and (2) "On the Formosan Spear-Fishes," by Kyosuke Hirasaka and Hiroshi Nakamura.

CHING CHANG WOO

Acta Geologica Taiwanica, vol. 1, no. 1 (April, 1947). Science Reports of the National Taiwan University, 1st ser., in co-operation with the Oceanographical Institute of Taiwan. Taipei, Taiwan, China: National Taiwan University.

The *Acta Geologica Taiwanica* is sponsored by the Department of Geology of the National Taiwan University, under the chairmanship of Dr. Ting Ying H. Ma. The first issue of this new publication contains three papers: (1) "Amount of Emergence of Eastern Asia during the Middle Pleistocene Time," by Ting Ying H. Ma; (2) "A Permian Cephalopod Faunule from Chêchiang Province, China," by Ichirô Hayasaka; and (3) "Fossil Species of *Clypeaster* from Taiwan," by Ichirô Hayasaka and Akira Morishita.

CHING CHANG WOO

Historical Geology. By RUSSELL C. HUSSEY. 2d ed. New York and London: McGraw-Hill Book Co., Inc., 1947. Pp. ix+465; figs. 322. $3.75.

Historical geology is so large a subject and includes so many fields of study that any treatment of it in a single text must necessarily be elementary. Professor Hussey was apparently well aware of this when he wrote *Historical Geology*.

The introductory chapters cover such topics as the origin of the earth, the geologic time scale, superposition, stratigraphic correlation, and paleogeography. The closing chapter deals with the geologic history of man. The body of the text is subdivided on the basis of geological time. The Archeozoic and Proterozoic eras are covered in a chapter each. The Paleozoic and Mesozoic eras are treated on the basis of one chapter for each period, and the Cenozoic era is allotted one chapter for the Tertiary and one for the Pleistocene.

The new edition is somewhat shortened by the omission of the appendix, which contained a short résumé of the phyla of the plant and animal kingdoms. The remainder of the text has been changed in several minor respects. The order of the introductory chapters has been reversed: the chapters on evolution and paleontology are placed after those on physical geology and stratigraphy. A few illustrations have been added. The most noteworthy changes in the new edition are the omission of the tables of formations for the various periods and the addition of generalized sections for several of the periods.

In several cases the author presents but one side of questions which most geologists consider far from settled. An example of this is the discussion of Gondwana Land, a hypothetical land bridge which crosses a present ocean basin. Notwithstanding the possible affinities of geographically widely separated fossil forms, the physical evidence does not support the possibility of land bridges of this type. The conclusion of many of the workers in animal geography, such as R. Hesse, W. C. Allee, and W. D. Mathew, with regard to widely separated but closely related forms is that such land bridges are not necessary to account for the distribution of these animals.

The remarks on *Seymouria* are misleading because the position of this animal in classification is far from settled. In this and other cases the artificiality of classification and its primary use for convenience only should be pointed out to the reader.

The fact that controversy exists on these as well as other topics discussed by the author should be noted in the text. It is the opinion of the reviewer that unsettled questions should be pointed out to the student who is seeking an introduction to any science, so that he may become aware of the challenge and fascination that lie in exploring the unknown.

ROBERT L. MILLER

Fundamentals of Physical Science. By KONRAD BATES KRAUSKOPF. 2d ed. New York: McGraw-Hill Book Co., Inc., 1948. Pp. 676; figs. 330; pls. 1. $4.50.

This is one of the best—if not the best—textbooks available today for use in a general-education course in the physical sciences. It is of particular interest to geologists because of its emphasis on their science and on those parts of physics, chemistry, and astronomy necessary to a geologist.

In the Preface, Professor Krauskopf writes that his purpose is to deepen and enlarge the reader's interest in science by stressing the *methods* of scientific reasoning rather than their results. These few words give a formal solution, at least, to the problem of general education in science—the problem which is central in such a text as this one or in teaching a general-education course—of how best to give some broad understanding of science in a limited space or in a limited time.

The dangers are, on the one hand, of presenting too shallow a survey of the subject and, on the other hand, of going so deeply into one part that no picture of the whole is presented. Or, again, if insufficient attention is given to detailed facts, a complete discussion of methods becomes meaningless, while if the facts are heavily stressed, as in the traditional elementary textbook, then there is no time left to discuss the methods; in which case the casual student, upon finishing the book or course and passing the examination, promptly forgets the facts and is no better off than if he had never studied the subject.

Professor Krauskopf starts off by avoiding these dangers with a facility that is little short of genius: in the first hundred pages he covers positional astronomy, the elementary mathematics necessary to explain mechanics, mechanics itself and its application to the solar system—all in a brief, interesting, and understandable manner. The style can be paraphrased: "Here is what we see; how can we understand it?"

In Part II, "Matter and Energy," a well-integrated sequence of eight chapters carries the reader through kinetic molecular theory and chemical reactions to a fair understanding of the periodic table of the elements, based on the gross characteristics of matter. But in Part III, "The Structure of Matter," the style deteriorates somewhat toward the dogmatic, possibly because of the complex subject matter (atomic physics and physical chemistry) or possibly because the author feels that the reader is by now captivated and neither needs nor wants the naïve approach.

Chapter xxi, "The Atomic Nucleus," apparently re-written for this second edition to draw upon popular interest in the atomic bomb, is unfortunately sandwiched into the middle of a discussion which should lead without impediment to an explanation of chemical phenomena in terms of electric forces. The resulting confusion is most evident in the treatment of light: an introductory chapter, "Light Waves," following hard on "Electric Currents," is incomplete and is separated by two chapters ("X-Rays and Radioactivity" and "The Atomic Nucleus") from its sequel, "Radiation." After this oscillation the reader is plunged, all unawares, back into "Subatomic Chemistry."

Part IV, a section of one hundred pages, entitled "Fundamental Processes," covers ionic reactions, chemical equilibrium, carbon and silicon compounds, and colloids. This undisguised inflation of chemistry at the expense of physics would normally be undesirable in a well-balanced survey of the physical sciences. In this case it serves to support the extensive Part V, one hundred and sixty pages on the "Biography of the Earth."

The first chapter of Part V disposes of seismology somewhat too briefly in a mere three pages and includes no application of the earlier study of wave motion. The chapter on meteorology is a clear, elementary presentation, although the discussion of the general circulation of the atmosphere is not too convincing. The following chapters are entitled, "Rocks and Minerals," "Erosion and Sedimentation," "Volcanism and Diastrophism," "The Law of Uniform Change," "Interpreting the Rock Record," and "Earth History." The author is to be congratulated on fitting together so much so effectively in so small a space. The methods of the geologist are clearly stated, and some of the philosophy of the science is discussed in the chapter on uniformitarianism.

Following this, the last, sixty-page section on astrophysics could scarcely avoid being anticlimactic. It is unfortunate that this section is so far separated from the chapters on light and that the treatment of Planck's law, Doppler effect, and double stars is not more quantitative. The existence of interstellar matter is completely ignored.

It would be possible to criticize many other omissions, of course, but a more basic criticism is that *Fundamentals of Physical Science* contains *too much* material for a sound, one-year course. The author himself suggests omitting several chapters in chemistry and geology, but this leaves the problem of supplying the connecting thread of the argument in the hands of the instructor. With this book as a text it will be appropriate to use geology as that thread; to adopt as the final goal in the student's mind the understanding of all the chemistry, physics, and astronomy which are necessary to understanding geology, as well as the geology itself.

The importance of books like this one can scarcely be overemphasized. It is true that the methods of science can be illustrated by the study of one science only—such as physics—but only an integrated picture of the physical world derived from the four sciences—astronomy, physics, chemistry, and geology—can show that *all* our physical experience fits into the pattern of scientific thought. Although it is not the final word on the subject, Krauskopf's *Fundamentals of Physical Science* represents a great advance in the integration of the physical sciences for elementary or nonspecialist readers.

THORNTON PAGE

Mineral Resources of Colorado. Prepared under the supervision of JOHN W. VANDERWILT. Denver: Colorado Mineral Resources Board, 1947. Pp. 547; pls. 34; figs. 10. $2.50.

This volume is divided into three parts: Part 1, "Metals, Nonmetals, and Fuels," by John W. Vanderwilt; Part 2, "Summaries of Mining Districts and Mineral Deposits," prepared by the United States Geological Survey under the general supervision of W. S. Burbank; and Part 3, "Investigations of Strategic Mineral Resources," by W. M. Traver, Jr., of the United States Bureau of Mines.

Part 1 contains a brief description of the general geology of Colorado, followed by a review, by counties, of the mineral production of the state. Numerous tables on production of minerals are included, as well as the principal references to each area. Vanderwilt has gathered together a huge amount of data, some of which was not in the literature, which should serve as a valuable reference.

Part 2 should be of interest to the mining geologist, teacher, and student. Emphasis has been placed on districts and, in some cases, metallogenetic provinces. Such a treatment makes the volume of much greater value than if it had been merely a geologic summary of the important mines of the state. It is to be regretted that some of the chief producing areas, such as Climax and Aspen, are not included in Part 2 and receive only the briefest treatment by Vanderwilt in Part 1. This may be due to the fact that these areas were recently reviewed in guidebooks of Colorado geology. In most cases the geologist most familiar with an area contributed the summary to this volume. The bulk of the work was done originally by the United States Geological Survey in co-operation with Colorado organizations.

Part 3 is brief (26 pp.) and consists principally of a listing of the mines examined by the United States Bureau of Mines as part of their strategic and critical minerals investigations. The sixteen exploratory projects of the bureau in Colorado are described.

This volume is primarily a review of Colorado mineral production and the geology of the important mining districts, including pertinent geologic literature. To the student and teacher, however, the value of the book is considerably increased by its emphasis in Part 2 on mining districts and, in some cases, on metallogenetic provinces. The wealth of descriptive material on Colorado mines has been digested and summarized to present the general geologic picture of the more important mining districts.

A. F. HAGNER

Symposium of Information Relative to Uses of Aerial Photographs by Geologists. Compiled by H. T. U. SMITH. (*Photogrammetric Engineering*, vol. 13, no. 4.) 1947.

The following papers are included, and after each a note is appended to indicate the contents. Much of the material has been printed previously, but some articles are new contributions. The symposium is particularly valuable inasmuch as it brings under one cover the important aspects of the use of aerial photographs in geology.

"Reconnaissance Mapping from Oblique Aerial Photographs without Ground Control." By JOHN L. RICH.

A method presented without use of Canadian grid or spot-location, but by use of transparent templates, so that position of faults, contacts, or position of strike and dip determinations can be approximately shown on the map.

"Geologic Interpretation of Trimetrigon Photographs, Northern Alaska." By SHERMAN A. WENGERD.

Recounts the procedure of geologic exploration of northern Alaska and the use and interpretation of trimetrigon photographs and the compilation of a structure map of the region from them.

"The Use of Aerial Photographs in Soil Mapping." By MARK BALDWIN, HOWARD M. SMITH, and HOWARD W. WHITLOCK.

General review of material printed in previous articles on the subject.

"Aerial Photographs in the Geological Study of Shore Features and Processes." By ROBERT S. DIETZ.

Reviews the kinds of shore features vividly shown on aerial photos, the use of photos in the study of waves, shore processes, and shore-line changes.

"Geological Mapping of the Ross Lake Area, Using Air Photographs." By Y. O. FORTIER.

The use of photos when no base map is available is described.

"Air Photographs in Geologic Mapping of Cordilleran Region, Western Canada." By A. H. LANG.

Remarks on preliminary office examination, planning traverses, plotting outcrops, obtaining rough idea of rocks, plotting geological boundaries, interpreting structure, extrapolating between traverses, physiographic data, and use in office compilation.

"Aerial Photographs as Aids in Stratigraphic Studies." By R. C. MOORE.

Detailed picture of strata is invaluable to guide stratigrapher to best outcrops and to help him study facies change.

"Aerial Photographs in Geology." By WILLIAM C. PUTNAM.

Review of general applications to geologic studies.

"Geologic Applications of Oblique Photography." By JOHN L. RICH.

Emphasizes use of obliques stereoscopically and explains briefly the use of obliques in map-making, how to determine dip, strike, and structure contours, and how to make a controlled aerial traverse.

"Advances in the Use of Air Survey by Mining Geologists." By G. W. ROONEY and W. S. LEVINGS.

Outlines use of photos in mining surveys in Canada, particularly the Beatty-Munro area, Ontario, and describes a course in aerial photos at the Colorado School of Mines.

"Use of Aerial Photographs in Glacial Geology." By F. T. THWAITES.

Emphasizes ability to study form of glacial deposits with aerial photos, which is of much value in mapping them.

"Multiplex Compilation of Geologic Maps." By JOHN D. STRABELL, JR.

An important contribution explaining the use of the multiplex in transferring geological data plotted on aerial photos to a base map, where considerable accuracy is required, as in mining districts.

"Aerial Photos in Geological Training." By H. T. U. SMITH.

Emphasizes advantages of aerial photos in teaching and as a new important technique in geologic mapping and research.

"A Method for Preparing Stereoscopic Aerial Photographs for Reproduction in Quantity." By JOHN L. RICH.

Where numerous copies of stereopairs are needed for classroom study with a lens-type stereoscope, a simple method of copying and printing standard 7×9 or 9×9 photos to the scale demanded for full coverage by the lens-type stereoscope is described.

A. J. EARDLEY

The Nature and Origin of Limestone Porosity. By Richard B. Hohlt. (*Colorado School of Mines Quarterly*, vol. 43, no. 4.) 1948. Pp. 48. $1.00.

Because of an increasing interest in the nature and origin of porosity in petroliferous limestones and dolomites, the author analyzes existing concepts and presents petrofabric data that bear on some of these concepts.

The first half of the paper is a discussion of primary porosity, "developed prior to or during lithification," and of secondary porosity, "developed subsequent to lithification." Because the economically important limestones are characterized by secondary porosity, primary porosity is only briefly outlined. Under secondary porosity "modifications of primary openings by solution action, jointing, fracturing and dolomitization" are considered.

In the latter half of the paper the author follows Bruno Sander, who first noted that the *C*-axes of calcite in limestones invariably lie in the plane of the bedding but with a random orientation within that plane, whereas dolomite crystals nearly always show a random orientation, although a few may show an orientation similar to that of limestone. Noting the diverse modes of origin of dolomite and the likelihood that dolomites with a preferred orientation are of primary origin, it is implied that limestones which have been dolomitized show a random orientation. The author concludes, other things being equal, that these newly formed dolomites ". . . may be expected to develop better solution porosity than a limestone . . ." due to this random orientation. The mechanism proposed for the formation of random orientation during dolomitization is that the newly formed dolomite crystals will orient themselves in response to zones of pressure release caused by the volume adjustment of Mg for Ca and selective leaching. In this study six petrofabric diagrams with corresponding photomicrographs of six different dolomites are included. However, these diagrams suggest a preferred orientation rather than the random orientation described. Five of the six diagrams show girdles containing point of maxima near the bedding plane, the sixth has a distinct girdle at a low angle to the bedding plane. There is little, if any, difference in the diagram by E. B. Knopf which was presented as an example of a "high degree of preferred orientation" of calcite and those described by the author as "random." On the whole, this study of porosity is stimulating, and the use of petrofabrics suggests a new approach to the basic problems of reservoir rocks.

F. A. Johnson, Jr.

Emanuel Kayser's Abriss der Geologie, Vol. 2: *Historische Geologie* ("Outline of Geology—Historical Geology"). By Roland Brinkmann. Stuttgart: Ferdinand Enke Verlag, 1948. Pp. vii+355; figs. 64; pls. 58. Unbound, M. 25; bound, M. 27.

For many decades, Kayser's textbook of geology has enjoyed a high reputation among European standard texts. Before the war the two-volume *Abriss* was supplemented by a four-volume *Lehrbuch*. Under extremely difficult conditions Brinkmann has succeeded in bringing the two-volume work up to date. (So far, the reviewer has not been able to obtain the first volume of this *Abriss*.) As Brinkmann states in the Preface, he has endeavored to supplant the dominantly stratigraphic presentation of historical geology in older editions by a closer elaboration of structural evolution of the earth's crust and paleogeography. The result is a most interesting, well-balanced, and clearly written representation of geologic history. The diverse, ever present mutual relations between areas of erosion and deposition, types of sediment, facies variations, volcanism, and crustal deformation are emphasized throughout the book, and the reader cannot but regret that the scope of the book prevented the author from giving much more detailed discussions.

After an introductory chapter on methods of measuring geologic time, the essential facts and interpretations of geologic history are presented. To compress the voluminous material, Brinkmann has added a great many new maps, showing paleogeographic elements, such as areas of contrasting facies, zones of regression, areas with related faunas or floras, shore lines of successive marine oscillations, marine-epicontinental versus marine-geosynclinal regions, isopach maps, partial reconstructions of climatic zones, distributions of particular faunas, etc. Other charts illustrate the stratigraphy of particularly interesting districts: pollen diagrams, progressive changes of selected families of organisms, climatic fluctuations, and sequences of earth events, similar to those employed by Von Bubnoff, Reed and Hollister, and others. As the book will be used for the most part by geologists in Europe, it is understandable that this continent receives more attention than others. Moreover, the availability of particularly complete data for Europe adds to the reliability of most of these charts.

Eight large, folded plates correlate the stratigraphy of the Ordovician through Carboniferous and Triassic through Tertiary systems. Other periods are represented on smaller charts. They have all been worked out with great care, and much non-European literature up to the beginning of the war has been incorporated. At the end of each chapter are found selected references. They are well chosen, but the reviewer would like to see at least twice as many. If they were printed in small type and in consecutive lines, they would not increase the length of the chapters but would be of great aid as short cuts in our exceedingly scattered literature.

The book ends with a ten-page review of organic evolution, as well as the inorganic evolution of the earth's crust from the pre-Cambrian to the present. The book has a comprehensive index of 36 pages and is practically free from typographic errors. Reproduction of fossil plates and typography are very satisfactory, and the publication of this book is another gratifying sign that standards of technical European books remain high, despite the numerous complications of the postwar situation.

R. B.

Journal of Petroleum Technology, vol. 1, no. 1 (January, 1949). Dallas, Texas: American Institute of Mining and Metallurgical Engineers. $8.00 in Western Hemisphere, $9.00 elsewhere ($4.00 to A.I.M.E. members).

This journal is the official monthly publication of the Petroleum Branch of the American Institute of Mining and Metallurgical Engineers. It, together with its two sister-journals, *Mining Engineering* and *Journal of Metals*, supersedes the old *Mining and Metallurgy* as well as the four "technologies" (*Petroleum Technology, Mining Technology, Coal Technology,* and *Metals Technology*).

In the *Journal of Petroleum Technology* are found those items of special interest to petroleum geologists and engineers, items of Institute-wide character (which appear in all three of the new journals of the Institute), and the technical papers on petroleum geology and engineering which formerly appeared only in *Petroleum Technology.*

The segregation of subject matter into these three specialist journals should make each a more valuable publication for those who are specialists in the three respective fields. The style and format of the first issue of the *Journal of Petroleum Technology* is strongly reminiscent of the parent-journal from which it comes.

F. J. P.

International Rules of Botanical Nomenclature. Compiled by H. W. Camp, H. W. Rickett, and C. A. Weatherby. Waltham, Mass.: Chronica Botanica Co.; New York: Stechert-Hafner, Inc., 1947. Pp. 120. $3.50.

This volume is the result of action of the Committee on Nomenclature of the American Society of Plant Taxonomists. It represents a compilation "of what already has been acted upon favorably or authorized at official sessions of the Subsection for Nomenclature of the more recent International Botanical Congresses, together with certain items pertinent to the work of plant taxonomists and which legitimately should be included in such a volume."

As stated by the compilers, the volume is not to be construed as an official edition. It has been taken from original sources, and every effort has been made to avoid the introduction of errors and to interpret certain ambiguities as carefully as possible. The format of the third (1935) edition of the *Rules of Botanical Nomenclature* is followed throughout much of the volume with, however, rearrangement in the Appendix containing the conserved generic names.

The purposes of the volume are set forth in a short preface from which the material of the first two paragraphs of this review have been quoted or paraphrased. The book is attractively composed, and the typography is excellent. A utilitarian index that adds greatly to the ease of use of the volume has been included. Botanists and paleobotanists should find the publication of great service in their taxonomic studies.

E. C. O.

Submarine Geology. By FRANCIS P. SHEPARD. New York: Harper & Bros., 1948. Pp. vii+348. $6.00.

Submarine Geology carries the reader from the coasts and beaches, across the submerged continental shelves, and down their slopes to oceanic depths. He gains a comprehensive view of submarine topography and structure, as well as of the makeup of the present-day sea bottom. The author has incorporated the most recent data from all sources which this branch of oceanography has produced in its rapid development of the past few years. Much of the information in the chapters on the dynamics of beaches, the classification of coasts, the sediments of the continental shelves and slopes, and submarine canyons, as well as of the deeper ocean basins, stems from his own investigations.

The stratigrapher will do well to consider the sedimentary evidence presented in chapter vi. To those accustomed to thinking of an orderly gradation, from coarse to fine material, in an offshore direction and of uniform deposition over large areas, the patchy distribution of the sediments on the shelf will come as a surprise. Furthermore, the lateral transitions from one type of sediment to another are commonly abrupt. However, it should be remembered that these sediments are largely deposited in major oceans and that, at present, no water bodies really comparable to the epeiric seas of the past are known.

The origin and structures of continental shelves and slopes has always provoked speculation, and several alternative explanations are presented by Shepard, who arrives at the conclusion that the evidence points to a complex origin. A valid distinction is drawn between shelves off glaciated and off nonglaciated regions on the basis of topography and bottom sediments.

Atolls, guyots, and submarine canyons are cited as possible evidence of deep, world-wide submergence following a period of emergence. The reader is left with little doubt that the canyon problem is still unsettled, in spite of the author's strongly stated position; but there can be no questioning of the facts presented, many of which were obtained from the author's own field experience. Indeed, the diametrically opposed arguments that have been evoked whenever the question of origin is discussed should indicate to the impartial observer that, even at the present writing, the problem is far from settled. Witness the controversy between Kuenen, now the chief advocate of erosion by the muddy suspension current, and Shepard within the last year. However, the authors of the many hypotheses proposed to explain these puzzling gorges have one thing in common. Each one appeals to factors decidedly out of the ordinary, set forth in one single, comprehensive theory. Perhaps this appeal to the cataclysmic may eventually prove justified; but in this connection, in view of the somewhat dogmatic arguments advanced, a few lines from *Troilus and Cressida*, brought to my attention by H. W. Menard, Jr., seem apt to the still skeptical:

Take but degree away, untune that string,
And hark what discord follows. Each thing meets
in mere oppugnancy. The bounded waters
Should lift their bosoms higher than the shores
And make a sop of all this solid globe.

Taken as a whole, *Submarine Geology* is an excellent exposition of the accumulated facts which are set forward without bias in the opening sections of each chapter. Shepard gives his preferred explanations of them, but there is nothing to prevent the questioning reader from placing his own interpretations upon them.

H. C. STETSON

The Permo-Triassic Formations. By R. L. SHERLOCK. London: Hutchison Scientific and Technical Publications, 1947. Pp. 367; figs. 15.

R. L. Sherlock has made a vigorous attack upon the problem of late Paleozoic and early Mesozoic time in this challenging treatment of the Permo-Triassic formations. In the concluding section of the book he recommends that recognition of the Permian system be abandoned in favor of association of the early Permian with the Carboniferous and the late Permian with the Triassic. While ingeniously argued and in many respects strongly documented, this proposal is likely to find few converts; for the inertia resulting from the weight of usage will be difficult to overcome. Although the proposal is stated in the closing section of the book, it is a theme which runs through many other parts and the framework upon which the book has been constructed. In some respects, however, this aspect of the work is less important than others, for divisions of geological time are, at best, arbitrary and are natural only in terms of the criteria which must be called upon to subdivide the continuum of time into workable units. For this reason recognition of the Permian system, or any other system, must depend in large part upon practicality and usage. Arguments similar to those advanced against acceptance of the Permian can be raised to a greater or lesser degree for essentially every geological system; and, in view of the historical development of the geological timetable, the nature of the rock record, and the criteria available for differentiation of various major rock units, this seems inevitable.

Part I of this volume is concerned with several items pertinent to a study of the Permo-Triassic—the origin of red beds, climatic conditions, geography, paleontology, and, perhaps somewhat out of place in this volume, economic products. Red beds are considered to be predominantly continental in origin and, in most instances at least, indicators of semiarid or arid climates at the time of deposition. Although the author does not believe them to be infallible indicators of climate, he does cite red beds as supporting evidence of aridity. The rather impressive array of data which suggest that red beds may have originated under a wide variety of circumstances, whenever sufficient iron was present under appropriate oxidizing conditions prior to, during, or after deposition is not, perhaps, stressed sufficiently.

In the chapter dealing with "Climatic Conditions of the Permo-Trias," considerable space is devoted to the concept of "continental drift." In the course of the discussion the writer states: "We must conclude that the fact of 'Continental Drift' has been proved, although the mechanisms by which it was effected and the detailed results are still matters open for discussion." The considerations of climate are, of course flavored by this conclusion. The arguments carry over into the chapter on "Geography of the Permo-Trias." Aside from the paleogeographic picture, which differs sharply from that presented in most North American literature, and illuminating discussions of local Permian and Triassic conditions, some insight into the thinking of the author on the general problems of division of geological time is provided by this chapter. He does not admit the possibility of time intervals between periods, a conclusion certainly logical from a theoretical standpoint but most difficult to apply. Later in the book and bearing upon this point, he notes that the type Permian section in Russia is very incomplete and thus not suitable as a type section for a system. Completeness, of course, can be only a relative criterion, since no type section of any system records all the time of the period and some systems even lack well-defined type sections. Subsequently, however, the author uses as partial justification for the assignment of the late Permian to the Triassic the evidence of a "break" between lower and upper Permian beds. It is not entirely clear whether or not the faunal and floral discontinuities at this break are thought to imply a time interval. Sherlock's philosophy of the bases for division of geological time is not stated precisely in the volume, and efforts to reconstruct it lead to somewhat contradictory conclusions.

The chapter on "Paleontology" gives a short summary of the invertebrates, vertebrates, and plants of the two periods. For the purposes of this book the summary is satisfactory, although parts of it are rather out of date. Further discussion of the meaning of the various groups of organisms is given in later chapters. Two statements in the chapter are of particular interest. It is said that plants and animals are rarely fossilized in the same beds; hence correlation of floras and faunas are difficult. This is true so far as invertebrates and plants are concerned, but it is not the case for plants and vertebrates, which occur together under a wide variety of circumstances. The second statement is to the effect that in cases where the evidence from plants and that from animals is conflicting, it is better to give greater weight to the evidence from animals for practical reasons. This emphasizes the arbitrary nature of the criteria, for it is not infrequently true that the position of a major break in time will be determined quite differently if the plants or the animals are used alone. The dependency of a "natural" system upon the weighting of the criteria may, indeed, be very strong.

Part II deals with the "Stratigraphy of the Permo-Trias." In many respects this section, which comprises 157 pages, is the most important in the book. Treatment is by geographic units, and discussions, which draw heavily on the cited works of many writers, are moderately detailed. A compendium of information on the Permo-Trias is thus made available for the first time. The section presents the background for Section III on "The Classification of the Permo-Trias." The keynote to this section is set by the writer as follows: "The root of the problem of the Permian System lies in the fact that no one has succeeded in determining a Period of time that is not represented by either Carboniferous or Triassic strata. In other words, all strata that have been called Permian may belong equally well to either the Carboniferous or the Triassic Systems." The arguments in this section are detailed and become as involved as the first quoted sentence is confusing. Few will disagree with the statement that the transition from the Carboniferous to the Permian is not clearly defined at many places either physically or biologically or that it is difficult to delimit the base of the Triassic in many instances. On the other hand, it seems adequately demonstrated, although the author seems to question it, that rocks were deposited between the times designated by established usage, varied as this may be in details, as Carboniferous, on the one hand, and Triassic, on the other.

Throughout the section, evidence of the twofold nature of the Permian is stressed, and the association of the earlier part with the Carboniferous and the later part with the Triassic is emphasized. The faunal and floral evidence for separating the lower and upper Permian beds depends in part on conflicting opinions and is not convincing. Evidence for relating the lower Permian beds to those of the Carboniferous is, however, strong. It is evident that geologists are faced with a difficult problem in developing a logical and practical division of the late Paleozoic and the early Mesozoic. Sherlock's proposal offers a solution based on fairly strong evidence so far as the criteria admitted for consideration are concerned. How much actual gain would come from adoption of the scheme is open to question. The sequence of beds in a given area and correlation of physical and biological events in time throughout the world are of utmost importance to the geologist in his task of reconstructing the history of the world. Any generally accepted series of time units can supply a frame of reference for this work as long as it is practical. There is serious doubt that Sherlock's plan is more workable the world over than is the one now generally in use in which the Permian system is recognized.

E. C. O.

Physiography of the Canadian Cordillera, with Special Reference to the Area North of the Fifty-fifth Parallel. By H. S. Bosteck. (Canada Department of Mines and Resources, Geological Survey Mem. 247.) Ottawa, 1948. Pp. 106; pls. 32; colored map of physiographic subdivisions. $0.25.

As the result of extensive air photography by the United States Army Air Forces and the Royal Canadian Air Force, the author is able to present for the first time a unified physiography of the entire Cordilleran region in Canada. The area north of the fifty-fifth parallel is treated in detail, while that to the south, which is much better known, is described only as far as necessary to indicate regional relations.

Most of the report is given over to naming, outlining, and describing the larger physiographic units. Their relations to geologic structure and glaciation are considered, and in many areas extensive erosion surfaces are noted. The lack of geomorphic interpretations emphasizes the fact that here is a geologic frontier wide open to detailed investigations. The thirty-two airplane photos are one of the most interesting features of the report.

L. H.

COMMUNICATIONS AND ANNOUNCEMENTS

ANNOUNCEMENT OF NEW JOURNAL

With the aid of a grant from the Swedish government, the Swedish Geophysical Society has recently initiated publication of a quarterly journal, *Tellus*. This journal is to be a medium of publication for original contributions, survey articles, and discussions in the field of geophysical sciences. Each issue will contain 48 to 64 pages, with a page size of 179×250 mm. All articles in the first issue were in English, although articles in the French and German languages will also be accepted. The initial issue contained an article on the geochronology of the deep ocean bed by H. Pettersson, meteorological papers by T. Bergeron, E. Palmen, A. Nyberg, and C. G. Rossby, and a paper by O. Rybeck on recent ionospheric measurements.

The subscription price is $6.00 per year. Communications relating to manuscripts and subscriptions may be addressed to the Editor of *Tellus*, Fridhemsgatan 9, Stockholm, Sweden.

ERRATUM

In the article, "Oriented Lakes of Northern Alaska," by Robert F. Black and William L. Barksdale, vol. 57, no. 2, the captions under plate 1, *A*, and 1, *B*, are reversed.

The photographs from which plates 1–4 of this article were made are used through the courtesy of the United States Army Air Forces.

VOLUME 57 NUMBER 5

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CARLSBAD CAVERNS AND OTHER CAVES OF THE GUADALUPE BLOCK, NEW MEXICO[1]

J HARLEN BRETZ
University of Chicago

ABSTRACT

The limestone caves of this region were made under completely saturated conditions long before the present erosion cycle began and probably before the Pliocene Ogallala formation was deposited.

INTRODUCTION

In the combination of four unusual characteristics the group of caves in the Guadalupe Range of New Mexico appears to be unique. It contains one of the deepest known caves of phreatic origin (Carlsbad), it possesses caves with gypsum flowstone, it exhibits carbonate flowstone outcrops and stalactitic debris in the mantle rock above caves, and it yields evidence of former fillings of green—not red—clay. As in other cavernous regions described by the writer (Bretz, 1938, 1942), the caves of this group are older than the topography. They are older than the great Capitan reef scarp in which and behind which they occur, older than the broad Pecos lowland east of the scarp, older than the latest uplift of the Guadalupe block, and probably older than the peneplain across the summit of the block.

Two of the seven caves studied in detail are partly or wholly in the Capitan reef rock, four are in the nearly flat-lying rock of the rugged summit country west of the scarp, and one is in the flank of a minor anticlinal fold well down the northern slope of the block. There are fifteen to twenty other caves in the region, some difficult and time-consuming to reach and all reported to be minor affairs.

The Guadalupe Mountain block (fig. 1) is triangular in ground plan and is tilted up along its southwestern margin. From an indefinite northern margin, possessing broad lowlands of 3,500–4,500 feet A.T. among its hills, in the latitude of Carlsbad, the general surface rises between two southward-converging lines of cliffs to culminate in about 35 miles as a great prow of 8,751 feet A.T., 4,000 feet above a salt flat to the southwest and some 3,000 feet above the country to the southeast. The western cliffs are largely fault-determined. The eastern scarp is commonly described as essentially the seaward slope of a great Permian reef, now standing in relief because of erosional removal of weaker sediments to the southeast.

The tilted triangular upland itself is

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[1] Manuscript received February 22, 1949.

Linlithgow Library.
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somewhat warped and faulted, and a considerable part of the western side is dropped in a series of narrow faulted slices known as the "Brokeoff Mountains," lying parallel with the main western cliff. Almost no drainage leaves the Guadalupe upland for this much-faulted country and the enclosed salt basin beyond. Most streams flow eastward or northeastward and escape by way of the block's lowest corner, where the reef pitches underground and its scarp disappears. There they enter the Pecos lowland.

A limestone gravel-covered plain at the foot of the reef scarp extends northeastward from the Texas–New Mexico line, becoming a part of the Pecos lowland nearer Carlsbad, where the gradually lowering scarp disappears. Close to the foot of the scarp the plain consists of imperfect alluvial fans, but most of it is a series of terraces dissected by Black River and its tributaries from the moun-

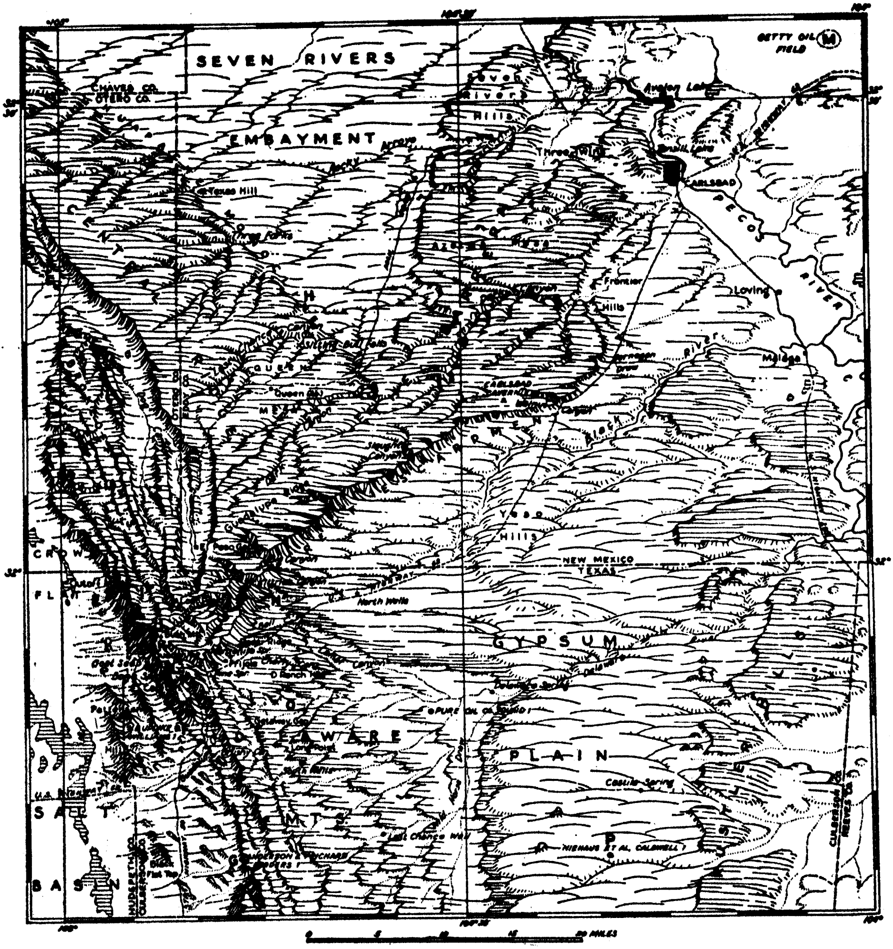

FIG. 1.—Map of the Guadalupe block and environs (from King, 1942)

tains and interrupted by collapse sinks. Black River low-water discharge vanishes under and reappears from beneath this fill several times in a stretch of a few miles. The subsidences and subterranean river courses are doubtless due to solution in underlying Permian fore-reef evaporites, chiefly gypsum.

In few places have geologists found more intertonguing and lensing and lateral intergrading of stratigraphic units than in the Guadalupe Mountains and the Pecos lowland and Delaware Mountains to the southeast and east. Most of the caves, however, are in the Capitan reef rock or its intertonguing back-reef correlative, the Carlsbad limestone. Their entrances are high in the spurs or walls of ravines cut by Pecos tributaries crossing the reef front. Carlsbad, by far the largest and deepest, is entered from almost the very brink of the scarp summit.

CARLSBAD CAVERNS

The known vertical range in this cave, from the ponor entrance on the ridge top to the surface of an unsounded pool in the lowest chamber, is 1,025 feet.[2] The entrance (4,350 feet A.T.) is 300 feet above the bottom of Walnut Canyon, 3,000 feet distant to the north, and about 550 feet above the base of the scarp, 3,500 feet away in the opposite direction. Thus the cave, extending down to 3,325 feet A.T., is an air-filled cavity, 725 feet below the bottom of Walnut Canyon and 475 feet below the upper limit of the gravel deposit at the foot of the scarp. This gravel plain, descending about 150 feet in the first mile away from the scarp and more gently with increasing distance, finally reaches the altitude of the cave bottom, 10 miles or more from the scarp.

The location of Carlsbad Caverns, more than 1,000 feet deep in a ridge 1¼ miles wide at the base and 400–600 feet high[3] (fig. 2), definitely disposes of the idea that the cave is a product of ground-water work during the present cycle of erosion. The cave system is older than the scarp. There is good evidence that the system once extended up into rock now eroded from the ridge summit and scarp face. The evidence is found in Chimney Cave and some associated topographic sags on the upper slopes of the scarp near Carlsbad's entrance. Chimney Cave is simply an unroofed vertical slot, reported to be 150 feet deep but to show no marked horizontal expansion. The opening is about 125 feet lower than that of Carlsbad and is on the top of one of the minor ridges between the short ravines scarifying the scarp face. The ridge east of Chimney Cave has a shallow ponor on its western slope, and, farther along the line between Chimney and the ponor, the crest of this ridge is made up of huge blocks of reef rock, lying at all angles, out of harmony with the otherwise uniform reef-bedding but displaced only by subsidence, not by having gravitated down along the slope. Still farther east along this line and across another ravine is a second shallow ponor, margined by big tilted blocks. Two or three other sunken tracts, 100–200 feet across, lie near by, likewise with displaced blocks that dip radially inward and also with flowstone fragments and a flowstone-cemented breccia in place (see pl. 1, *A*). Two good specimens of broken stalactites were found in these sags. One of the sags is reported to lie directly above the Jumping Off Place of the cavern.

Further evidence that Carlsbad was

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[2] Revision of figures supplied by the National Park Service and published by the author in 1942.

[3] The ponor entrance is 100 feet below the ridge summit.

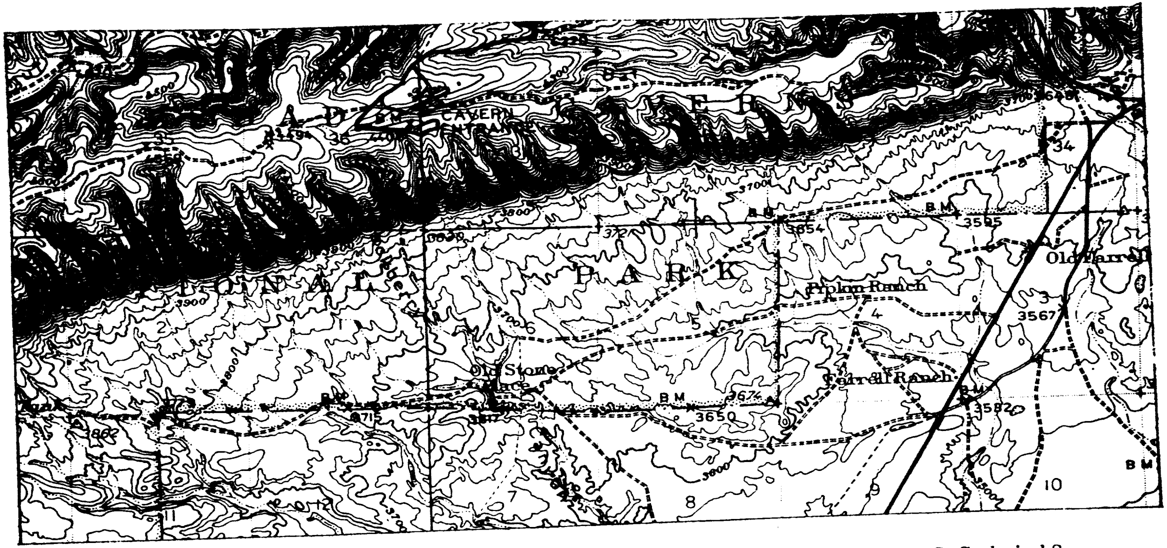

FIG. 2.—Map of environs of Carlsbad Caverns. From Carlsbad Caverns East topographic map, U.S. Geological Survey

made below a higher land surface than the present ridge summit is encountered immediately on entering the cave. The ponor collapse is asymmetrical, an uncollapsed part of cave wall and ceiling now being exposed to daylight. Tubes and pockets of solutional origin exist in the walls not more than 30–50 feet below the ground surface. They belong to that large category of minor cavities in cave walls and ceilings which record differentially greater local yielding to solution rather than locally greater vigor of attack on uniformly soluble rock (Bretz, 1942). Spongework belongs to the same category, and Carlsbad's magnificent display of spongework is to be seen from the highest to the lowest levels of the cave. Only below the water table can such solutional yielding occur on all bounding surfaces of a cave.

The argument for phreatic origin of Carlsbad also includes the joint-controlled network well shown by the cave map (fig. 3). Solutional enlargement and extensive collapse have so modified some chambers that, from inspection alone, one hardly recognizes the original joint control. But the great system has all gradations from narrow vertical slots to the Big Room, almost all elongated north-south or east-west, in conformity to the joint system. Particularly instructive is the Left Hand Tunnel, a narrow route to a great hole, the lowest place in the cave. The Tunnel is a tight, vertical slot quite unsuited for tourist inspection. In places its controlling joint is multiple, and the passage becomes four closely spaced slots with partial or complete partitions. The spongework of the Tunnel is unsurpassed in any cave that the writer has ever seen.

In three separate places Carlsbad possesses a record of vadose stream water whose flow was subsequent to the phreatic making of the cavern. One of them is a continuous stream channel, at least 1,500 feet long, in the Lower Chambers. It is situated so far down in the system that, if the former cave stream emerged at the foot of the scarp, there could have been no gravel fill within 200 feet of the altitude of the one now banked against that scarp.

An alternative to this suggestion is that an unknown subterranean route, miles in length, carried its water to the Pecos Valley. The river now is about 25 miles distant and 500 feet lower. The cave stream's gradient, by aneroid, was nearly 90 feet per mile. This must have been decreased to 20 feet per mile for this suggested long underground course, unless the valley was considerably deeper than it is today. Perhaps the extensive solution in evaporites of the Permian White Horse, Castile, Saludo, and Rustler deposits underlying the Pecos Valley provided opportunity for wholly subterranean discharge well over into Texas. It is difficult to dispose of the indicated volume and presumed load by infiltration through the gravel of the foreland, and a deep stage of Pecos drainage, either surface or subterranean, seems required.

This free-surface stream followed the floor of such phreatic passages as provided gradient for its needs. Direction of flow, recorded by gradient and by imbrication of cobbles, was south, then west, then *north*, away from the scarp! Obviously, it was not determined by a water-table slope and was not escaping very successfully from under the reef scarp. It found the floor of the chambers that it used already covered with silt, and it cut a channel in the silt from 3 to 4 feet deep at the accessible upstream end and from 16 to 18 feet deep at the lower accessible end. Gravel bars indicate a flood depth of 6 feet. The stream did not operate very long, for it failed to modify the limestone walls with the horizontal undercuts and meander niches that many vadose cave streams have made and are making on phreatic chamber walls.

Another vadose stream record exists off the northeast corner of the Lunch Room Chamber. The passage is low, with limestone gravel and later flowstone too near the ceiling to allow a traverse adequate for determining gradient. The cobbles, fairly well worn and ranging in diameter up to 16 and 18 inches, are im-

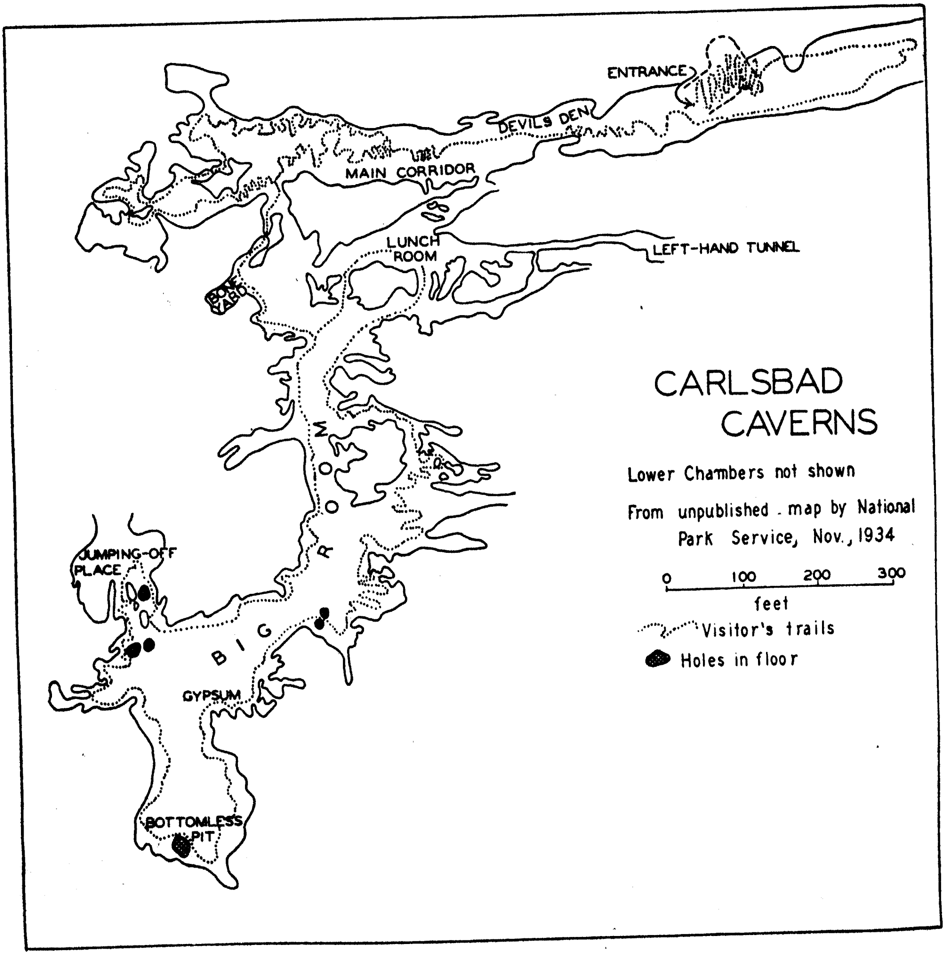

FIG. 3.—Map of main part of Carlsbad Caverns. After National Park Service unpublished map

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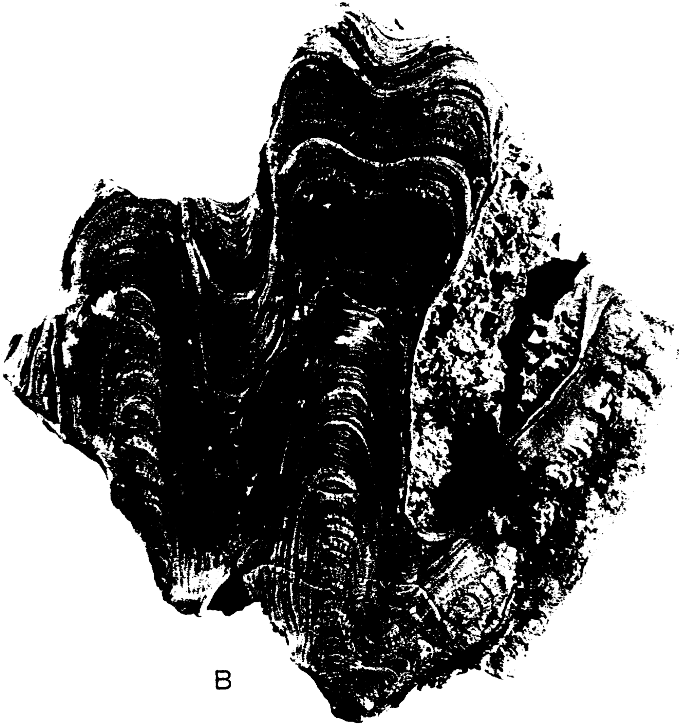
PLATE 1

A, Blocked cave opening on slope of reef scarp near Chimney Cave. L. Horberg, photo.
B, Flowstone from valley slope near entrance to Hidden Cave.

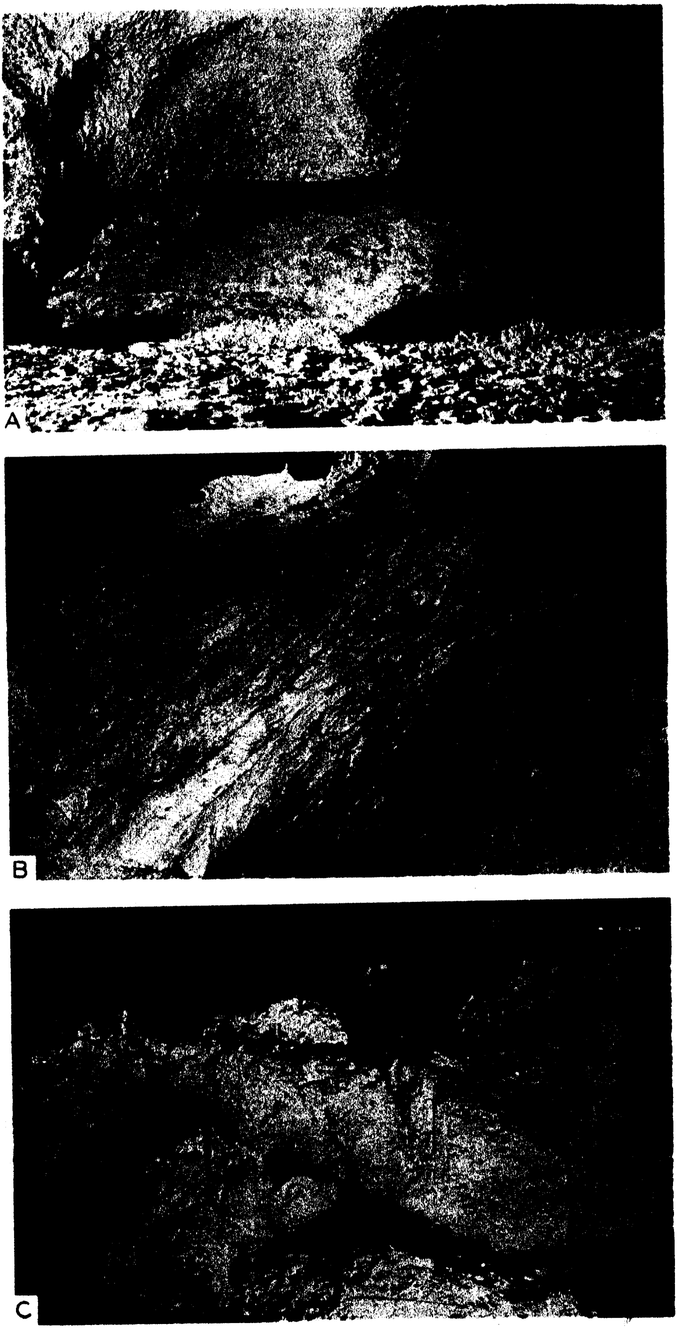
A

Blocked cave opening



B

Specimen of flowstone



Carlsbad Caverns features

bricated to show that the stream entered the much larger chamber from the north. Obviously, the Lunch Room Chamber (it seats 1,000 people) was already in existence when the vadose stream took origin.

The original floor of the Lunch Room had some deep slots or holes, now filled and covered with concrete.[4] Probably this vadose stream plunged down such a hole, to become, 90 feet lower, the stream of the Lower Chambers. Fallen rock and dripstone deposits prevent following the deeper portion upstream to the foot of the postulated fall.

Apparently this segmented stream of two portions succeeded an earlier and smaller vadose stream along the same routes, the record being the silt terraces (pl. 2, *A*) along both routes and also at the entrance to the Left Hand Tunnel from the Lunch Room. The deposit is well-sorted, laminated fine sand and silt, with no intercalated pebble layers. The terraces constitute flat-topped and accordant remnants, now separated because of the channeling of the cobble-carrying successor and, in the Left Hand Tunnel entrance, because of slump-pit development there. The change from a small, silt-depositing stream to a large, cobble-transporting one came abruptly. The earlier stream may well have been born of seepage water from the ridge above, but the later discharge can best be explained as an underground detour of some surface stream, which, in downcutting, discovered a part of the deep phreatic system and utilized it for a time. This is no improbable procedure; for other caves are known where surface streams, having done this very thing in the past, later returned to their surface valleys and deepened them well below any possible future reuse of the cave route. Walnut Canyon bottom, $\frac{1}{2}$ mile or so distant, is still 300–400 feet higher than the Lunch Room floor and probably was yet higher at the time that the cave stream functioned. Although, in flash flood, it is a considerably larger stream than the underground channel could care for, the suggested explanation can be adapted by detouring only part of the stream because of a constricted entrance route.

Stalagmites, coming to light as silt slumps down at the Left Hand Tunnel entrance, prove that the first vadose stream did not begin functioning for some time after the cave was drained.

A third deposit of stream gravel, apparently unrelated to the two already discussed, is near the top of the descent to Devils Den. It consists of limestone pebble gravel and fills phreatic wall pockets of a narrow, channel-like passageway lateral to the great main corridor. A few pebbles of green clay in the deposit were derived from near-by remnants of that clay in other pockets.

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[4] "Beneath the center of the Big Room floor is a maze of phreatic passage ways at a level intermediate between the Big Room and the Lower Cave. This consists of a network of fairly long tubes at right angles to each other and contains beautiful spongework. The presence of this intricate system along the two major joint zones gives further evidence that the Big Room was once a network of phreatic passage ways arranged both horizontally and vertically, and that continued solution and collapse have opened it into the present huge chamber" (Bennett Gale, personal communication).

PLATE 2

A, Silt terrace and cobble-covered channel floor, Lower Chambers, Carlsbad Caverns. B. Gale, photo.
B, Reef bedding in Capitan limestone, Bottomless Pit, Carlsbad Caverns. B. Gale, photo.
C, Drip solution holes in cave gypsum deposit, Big Room, Carlsbad Caverns. B. Gale, photo.

Green clay has been found in Mystery Room, in the Lower Chambers, and in several places in the Big Room. Its vertical range is almost as great as that of the cave. In both Mystery Room and the Big Room it has become oxidized or stained to a red or chocolate color on the surface and along laminations and desiccation cracks. Excavations for trail-making materials in the Big Room show no intercalated coarse material or secondary lime deposits. It may, therefore, be a phreatic clay deposit, made when the Guadalupe region was a peneplain, when differential hydrostatic pressures and circulations had disappeared because of the destruction of the prepeneplain relief.

There are, however, three arguments against this view: (1) In the Big Room, where much ceiling rock has fallen, some clay lies largely unburied, hence must postdate most of the collapse and should therefore be late vadose in age; (2) the ridge tops, even the broader, flatter summits, carry only a thin, interrupted soil mantle, and there is almost no clay in it; (3) the capacity of Carlsbad is so enormous that one hesitates to consider that the cavern has ever been clay-filled and then re-excavated, as have caves with similar phreatic-vadose history under humid climates. Further study may show that the clay in Carlsbad records more than one episode.

Remnants of an apparently once continuous deposit of white, massive, unstratified gypsum lie on the floor of the Big Room and in marginal chambers of the Lunch Room. The maximum exposed thickness is about 20 feet. It is younger than some carbonate flowstone, but great stalagmites sit on top of it and therefore are still younger. It seems that the gypsum can only be the consequence of local pooling during the vadose history, recording temporary conditions when sulfate alone was precipitated. A local source may have been the Permian gypsum both outcropping and known in wells close to the base of the near-by scarp and higher than these cave-chamber floors. Dripping water in many places has later dissolved vertical tubes in this gypsum (pl. 2, *C*). Some big blocks of the gypsum have lost 50 per cent of their bulk by such drilling. One hole is 12 feet long and only 3–4 inches in diameter. Elsewhere, carbonate flowstone and stalagmites have covered the gypsum.

NEW CAVE

About 9 miles along the scarp southwest from Carlsbad Caverns, the mouth of Slaughter Canyon constitutes a notch 1,300 feet deep in the reef front and a mile wide at the top (fig. 4). Two caves have been transected by this notch, one on each side. Each cave is about halfway up the canyon wall. The close proximity of the small Midnight Canyon, southwest of Slaughter Canyon notch, has made a sharp spur in the scarp front, and a saddle in the spur isolates a summit hill at the tip, less than 30 acres in area. New Cave lies under this hill, scarcely more than 150 feet below the level of the saddle.

Entrance to New involves descent of a long cave talus deposit whose debris has come from the canyon slope. The fan has spread a convex foot out into three subparallel chambers and has mounted up around the separating pillars or walls. The debris is composed of thin, flat slabs, dipping into the cave with the talus slope but obviously not derived from the local Capitan reef rock. It is Carlsbad waste and either came down from higher surfaces or was introduced by the stream when the canyon was only half as deep as now.

Two other transections of the cave system lie east of the enterable opening and at the same level. One may be traversed for about 50 feet to a jumping-off place, beyond which the spotlight shows considerable continuation. This cliffy slope with the three cave-chamber cross sections also possesses much slabby, cemented talus, the dip of the slabs being toward the wall, away from the canyon. Clearly, there were other old cave chambers here, now completely blocked with waste from the outside.

The total vertical range of New Cave is but little more than 100 feet. Most of the chambers are high and relatively narrow openings along three sets of joints, approximately north-south, north 40° east, and east-west. One group of six such chambers, varying in heights and lengths but all close-set and connected, possesses the potentiality of becoming one large chamber under sufficiently long-continued phreatic conditions.

The horizontal extent southward must bring the cave very close to the opposite

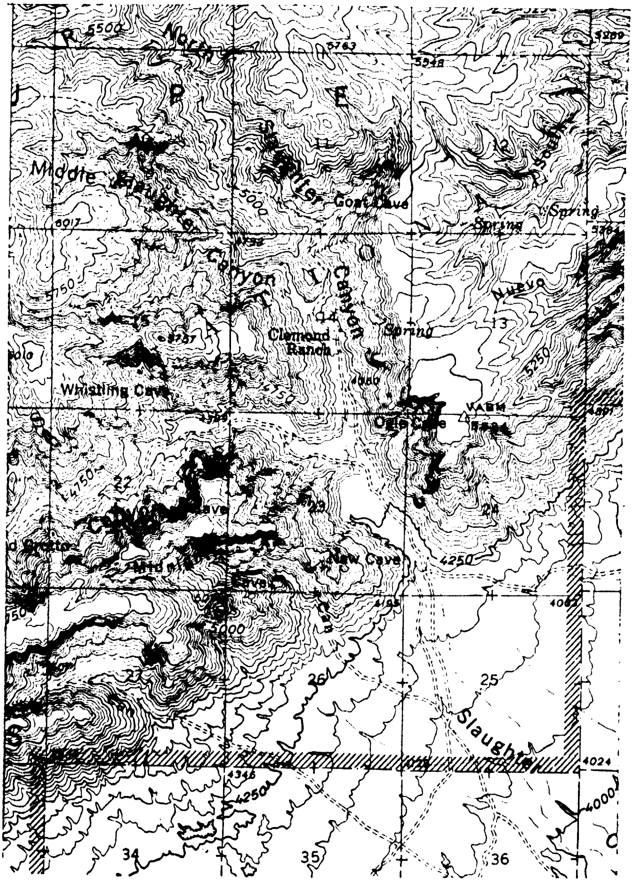

FIG. 4.—Map of Slaughter Canyon and environs. From Carlsbad Caverns West topographic map, U.S. Geological Survey.

side of the hill on the spur tip, under which New Cave lies. This idea is supported by the heavy flowstone deposits, which, sloping northward, block the cave at the south end.

No evidence of vadose water-work, other than the secondary deposits, was seen in New Cave. Although the canyon stream has cut away part of the cave, it apparently never found an underground traverse possible. The joint control of high, narrow, slotlike chambers, the bridgings across those slots, the windowed partitions, the ceiling and wall pockets, and the abundant spongework are as positive affirmations of phreatic origin as the cave's situation, high under the 30-acre tip of a spur carved from the reef scarp, is positive denial of vadose origin. New Cave, like Carlsbad, is older than the lowland east of the Capitan escarpment.

A feature of New Cave's dripstone, unique in the writer's experience, is solutional spongework or near-spongework in some of the stalagmites. Re-solution of stalagmites by drip water is not rare, but this spongework has been made, as has cave-wall spongework, under conditions of complete submergence. The best display is in the giant stalagmites near the entrance. They are old and dull and have suffered wastage mostly by the penetration of large irregular cavities on all sides. These cavities transect the concentric growth structure, producing a striking similarity to the grain of a block of carved wood. The outstanding exhibit is a compound stalagmite, 15 feet or so in diameter and 20 feet high. Differential solution has largely spared the top while eating away the sides, so that the form produced suggests a gigantic canopy umbrella or toadstool, the underside of the cap and the irregular stem showing the grain noted above. A few subordinate lower shelves or partial canopies are comparable to the cap. Later dripstone here and in many other places has covered some of these corrosional outlines. Unpocketed stalagmites must have grown later or have had such ravages entirely hidden by renewed growth.

Was this solutional submergence, after a vadose experience, an isolated affair in New Cave because of local and temporary conditions? Does it record a return to phreatic conditions along the whole scarp? Evidence presented later in this paper seems to make the second alternative as satisfactory a choice as the first.

Another interesting feature of New Cave's dripstone is that some of it is gypsum-coated and that some selenite crystals are ½ inch in diameter. The Christmas Tree stalagmite is a great compound form, with a snow-white coating of gypsum in which sparkle the larger crystal faces. No dripstone composed wholly of gypsum and no gypsum flowstone were seen in the cave.

COTTONWOOD CAVE

This cave is situated high in a divide, the Guadalupe Ridge (see fig. 1), which lies parallel with the scarp for 15 miles or so and some 4 miles back of it on the dissected oldland. The ridge presumably is an imperfectly reduced portion of the uplifted peneplain recognized in the summit uplands of the mountain block. The cave mouth is only 100 feet below, and ⅜ mile distant from the ridge crest at Dark Lookout, 6,850 A.T.

Cottonwood Cave is a simple linear chamber which extends for ¼ mile back under a minor spur among Black River headwater canyons. The detrital floor descends 225 feet in that distance, and its ceiling descends 150–200 feet. The controlling joint strikes north 15° west and in places shows in the ceiling as a solutionally widened slot with some surviving bridgings. The mouth of the cave is 500 feet above the canyon floor, $\frac{1}{8}$ mile distant.

The overlying ridge is so narrow that, if Cottonwood's orientation were 90° to its length instead of about 45°, the cave would completely perforate the ridge. With no other evidence of the character of the cave, the conclusion is inescapable that it antedates the canyoning. Surviving bridges high in the ceiling slot are positive evidence of its phreatic origin. The descent of the cave back into the ridge and under surfaces 100 feet higher than at the mouth is impossible for ground water, phreatic or vadose, to have made during the present cycle of erosion. Indeed, the cave comes so close to the trace of the peneplain that we seem almost required to date its phreatic origin as earlier in that erosion cycle.

Uncommon features of caves, but well shown in Cottonwood, are slicing and falling of nearly vertical slabs parallel to the wall. The partings by which future slices will become disengaged dip very steeply back into the walls on either side. Were the cave developed deep beneath a former land surface, these diverging cracks might be the record of incipient failure by shearing under heavily loaded roof rock.

Cottonwood has two groups of huge dripstone growths, one within reach of sunlight, the other about 300 feet farther back. Descent over some 50 feet of in sloping talus leads to the first and much more massive group. The thickness of the roof rock over this place is about 30 feet, and the volume of the secondary calcite deposits is so great that, if they have grown by contributions only from the existing roof, it should show evidences of this great loss. Yet the ground surface, 30 feet above, shows no trace of sinks or collapse or a generally carious condition. Nor could this calcium carbonate have been brought laterally from very far, for the ridge is narrow and steeply sloped. It is concluded that these marked accumulations were made before the heads of Black Canyon's two tributaries had reached the locality and, by their deep incision, had left the narrow ridge.

BLACK AND HIDDEN CAVES

Four more caves lie in this high country near Cottonwood, all close to the divide and none more than 2 miles from Dark Lookout. Two small unnamed ones are shown on the Carlsbad Caverns West topographic map under a spur ridge whose flat summit is essentially a peneplain remnant. Black Cave's entrance is 250 feet below this summit, near the bottom of Cave Canyon. The entrance to Hidden Cave is about halfway down the side of another small canyon, 500 feet deep at the cave.

Both caves, transected by ravine growth, have been nearly sealed up because of gravitative migration into them of hillside waste. Each cave has maintained an opening to the outside only because, as ravine widening has pushed the hillside back and shortened the cave, ceiling fracture at intersection with the hill slope has persistently continued and slope retreat has carried the opening higher and higher (fig. 5). To reach the true solutional cave, one descends a sloping chimney bounded on one side by fractured bedrock and on the other by in-sloping talus and flowstone.

Hillside waste below each opening contains fragments of dark-brown, laminated, sublustrous, slightly translucent travertine, identical with the weathered flowstone exposed at the entrances (pl. 1, *B*). One mass down the slope appears to be in place. Some of this flowstone is cementing material of a breccia. These materials are relics of the two caves' former extensions, now destroyed by later ravine-making. Because the elongation of Hidden is approximately 45° to the strike of its transecting valley slope, the travertine debris is traceable in a long diagonal down the slope. There can be no reasonable doubt that the canyons are later than the caves and that the lime which went into making the flowstone came from vanished roof rock.

of a small pool in which gypsum was deposited. The deposit is now disintegrating into a loose sandy detritus.

The strike of the cave is north 10° west, almost paralleling the ridge crest to the east and the valley bottom to the west. It is difficult to visualize either vadose or phreatic ground water following a course so related to its associated topography unless special conditions demanded. Joint control in a cave so close to a steep hillside seems quite inadequate, and no other cause is indicated.

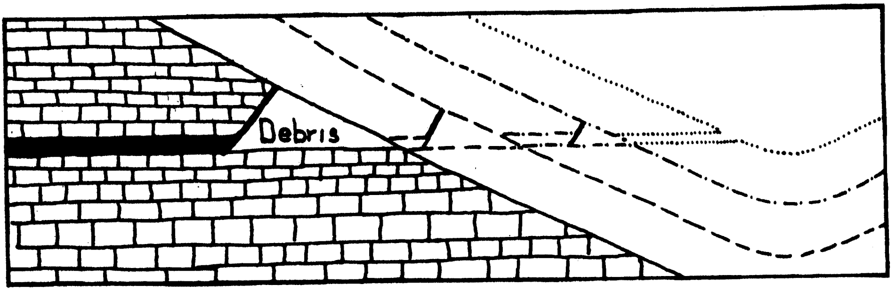

FIG. 5.—Diagram to show up-slope migration of entrance to Hidden Cave as the outside valley was enlarged.

Black Cave's chambers consist of three closely spaced, joint-determined openings, the separating partitions being scarcely thicker than the openings are wide. There is a short connecting cross-chamber. Walls converge upward to make steep-gabled ceilings, which repeatedly show solutional pocketings and groovings along the joint. Dripstone blockades terminate the open passages.

There has been much spalling from walls, some slices still unfallen but free from wall rock. This has destroyed original solutional outlines, has widened the passages from earlier slotlike proportions, and has made a fill of unknown depths.

In the wealth of secondary carbonate deposits in Black Cave, there is a record Black Cave, therefore, is confidently interpreted as a phreatic solution feature antedating the valley dissection of this plateau-like portion of the Guadalupes. It lies scarcely 250 feet below the peneplain trace.

Hidden Cave's ground plan involves six or seven connected, straight, narrow passages recording three joint sets, which strike north 10° west, north 27°–30° west, and north 80° east. The main north 10° west passage, where entered, is double, the separating partition hanging from the ceiling, its lower part completely dissolved away. The eastern of these two parallel passages terminates in a short distance with the joint showing in the end wall. It is essentially a huge wall pocket. At the south end of the traversable cave is a chimney slot which opens to daylight above and leads down 20–30 feet into an apparently east-west passage. Lack of sufficient rope prevented examination of this lower opening.

The imperfectly developed joint-controlled network plan is sufficient evidence that Hidden Cave is of phreatic origin. Vadose conditions arrived, and flowstone deposits were made before Dark Canyon's tributary heads had been extended back into the upland as far as the cave's location. Much of the original cave was later obliterated by valley erosion outside and dripstone deposition inside.

MUDGETT'S CAVE

The mouth of this cave is about halfway up the cliff of Dark Canyon in the last of the dozen or so large entrenched meanders called "Serpentine Bends." It is 300 feet above the stream bed and best reached from the divide between Walnut and Dark canyons. It is readily traversable for at least 750 feet back under the hill, its fairly straight horizontal course lying north 75° east, at right angles to the cliff face. The detrital floor slopes back into the hill, away from the canyon. Although there has been much wall failure, enough spongework remains to testify to a phreatic origin.

Three ravines have nearly isolated the hill above the cave, its small area and steep slopes above cave level being quite inadequate, even during earlier stages of the present erosion cycle, for gathering the ground water that made the cave.

There are two more caves in the vicinity of Mudgett's, each revealed by a chimney perforation at the surface but neither now possessing much horizontal extent underground. The chimneys have no associated sinks and appear to be collapses into chambers below or simple intersections of high slots by the erosional lowering of the surface, and not the products of descending vadose water.

One of the unnamed caves is about a mile south of Mudgett's, close to the bottom of Lechuguilla Canyon's minor dissection of the upland back of the scarp. Its chimney opening is determined by the intersection of two vertical joints. More high slots of the former cave appear to be recorded by two or three nearby outcrops of a dark, coarsely crystalline cave onyx or flowstone, between adjacent walls of Carlsbad limestone. The original laminations still show well, crossing the cleavages of the completely recrystallized material.

McKITTRICK CAVE

The only cave of this study lying well removed from the reef is McKittrick. It is about halfway up the eastern slope of a 250–300-foot hill near the northeastern corner of the Guadalupe block. The hill is on the apex of a definite anticlinal structure, and the cave is in strata dipping 10°–15° northeastward. Although no chamber is more than 20 feet high, the cave's enterable vertical range is all of 100 feet.

McKittrick Cave, early in its development, was a network of linear passages along two sets of joints, one in the strike, north 50° west, the other along the maximum dip. Widening of the intersecting joint slots of the original rectangular pattern has been so great that many of the once separating blocks, pillars, and partitions have been destroyed except near the upper end of the cave. But the ceiling throughout and the surviving attenuated pillars record the earlier joint control. Splendid dome cavities in the ceiling are also products of phreatic solution. Under later vadose conditions, the ceiling joints have determined lines of stalactites and stalagmites. Some partitions have been constructed by the growth of rows of coalesced columns. The cave is an excellent example of an intermediate stage between an early network and a late complete coalescence into a large room.

The only gypsum cave flowstone known to the writer covers a large portion of McKittrick's floor, extending from the highest to the lowest enterable portions and attaining a maximum exposed thickness of 5 feet. It originally had depositional terracings on its surface, the steepened fronts still recognizable. Partial destruction has been largely by later drip water which has drilled innumerable vertical holes in it, identical with those perforating Carlsbad Caverns' gypsum deposit in the Big Room. Partial obliteration has come from the growth of carbonate stalagmites on it. Some stalagmites, however, appear to be older and therefore to penetrate up through the sulfate flowstone.

The source for this gypsum is not the rock of the present hill. The cave is in the Yates limestone, well up toward the top of the Guadalupe series. The Ochoa rocks, stratigraphically about 200 feet higher and once overlying, are gypsiferous; so is the Yates or its correlative strata, about 5 miles to the west.[5] If the gypsum came from Ochoa strata, as seems probable, the cave became vadose while denudation lacked at least 200 feet of producing the present topography.

Like all Guadalupe caves—indeed, like almost all limestone and dolomite caves known to the writer—McKittrick Cave is in its decadence. It is filling with secondary precipitates and with collapse detritus from three collapsed places in its roof. There have been chambers of this cave above the present hillside surface, now completely erased, but with fragments of their flowstone and dripstone still lingering on the slope. Caves are destroyed under vadose conditions, rarely made.

AGE OF THE CAVES

Every Guadalupe cave examined has yielded what seems to be incontrovertible evidence that it antedates the present erosion cycle and therefore antedates the making of the existing reef scarp and the lowland east of it. Three of the caves are but little below the trace of the former oldland, which now survives only in isolated flattish summits. One cave records a vadose stream whose vigorous current found escape much below the level of the gravel flat abutting against the base of the reef scarp. For this escape, a postpeneplain Pecos Valley deeper than the present one seems required.

In addition to this stage of postpeneplain erosional history, another event in regional history must now be introduced, one that, for weeks of the field work, was difficult for Dr. Horberg and the writer to accept but which accumulating evidence during the work finally forced upon us. It is a depositional item; the existence of quartzose Rocky Mountain stream gravel, not alone in the Pecos lowland, where it had previously been reported (Meinzer, Renick, and Bryan, 1927; Fiedler and Nye, 1933; Morgan, 1938) but also on the summit of the reef scarp. Two significant occurrences will be described and their relation to cave history indicated.

The summit of the scarp between the deep notches cut by Rattlesnake and Slaughter canyons is an elongated, flattish surface descending gradually northeastward for 750 feet in a little less than 4 miles. Correlated with other comparable upland flats, both on the reef top and back of it, the flat appears to be a good record of the tilted peneplain (see fig. 4). Its highest or southwest end is

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[5] Oral information from Carl Branson.

5,500 feet A.T., nearly 1,500 feet above the upper edge of the piedmont limestone gravel deposit. This mesa-like upland carries many thousands of worn, siliceous stream pebbles and in one place shows an indurated remnant of the old quartzose gravel deposit *in situ*. The deposit occurs in an unroofed, surviving portion of a former cave!

The remnant, accompanied by several acres of scattered loose pebbles, lies at the head of Nuevo Canyon, between 5,425 and 5,475 feet A.T. Quartzite, chert, jasper, and vein quartz are all represented in what is probably the best display of these foreign pebbles seen in either the Guadalupe upland or the Pecos lowland.

Remnants of the cemented deposit range through 50 feet of altitude at the canyon head. The uppermost exposures show brown lime-sand strata associated with granule-gravel strata. Down along the ravined head of Nuevo for 200 feet and through a vertical range of 50 feet, remnants of a pebble conglomerate lie in cracks and cavities of the limestone. Most contacts are nearly vertical. In one is a slight overhang of the limestone wall.

This relationship would be an inadequate basis for the assertion that the quartzose conglomerate is a cave-filling almost on the brink of the 1,500-foot scarp. The former cave, however, is recorded by the largest and best display of weathered, recrystallized, laminated flowstone encountered during this study. Like the river gravel, it is distributed in loose pieces and is also in place, filling cracks and cavities in the limestone. As loose material, it occurs only in the immediate vicinity of flowstone and quartzose conglomerate *in situ*. Minimum dimensions of the former cave or group of solution cavities were about 200 feet horizontally and 50 feet vertically. The flowstone is considered good evidence of a once roofed condition and must be older than the summit flat.

A relationship sought by three pairs of presumably trained eyes was flowstone fragments in the conglomerate or siliceous pebbles in the flowstone or at least a flowstone-conglomerate contact. None was found. There are some large angular pieces of limestone embedded in flowstone, presumably fallen ceiling or wall rock. The conglomerate seems best considered as younger than the flowstone and surely is younger than the solution cavities which it fills. If the mesa top is a peneplain trace, the old cave was made still earlier and was destroyed in the final stages of that erosion cycle.

Further evidence that a scarp existed when the gravel was deposited is found along a poor trail traversing the second shallow draw on the scarp face, north of the mouth of Slaughter Canyon. Almost directly east of, and at least 400 feet below, the 5,524 bench mark on the upland's summit, two vertically elongated, dikelike bodies of indurated siliceous conglomerate crop out in the reef rock along the trail. One of them has much associated flowstone, but no contact of gravel and flowstone was found or fragments of either in the other. Whether or not these were solution slots, there apparently was a scarp face existing here at the time of gravel deposition, and it has retreated so little that pebbles still linger in its reentrant cracks.[6] If the indurated quartzose conglomerate being trenched by Black and Pecos rivers is a lower part of the same deposit recorded on the mesa summit, the scarp was even higher in prequartzose time, and, barring differential movement along the scarp, the deposit attained a thickness of at least 3,100 feet.

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[6] The only alternatives are to consider these as clastic dikes, with associated flowstone, whose pebbles migrated 400 feet down into open joint cracks, or as pebbles caught in these cracks as they slid down the present scarp face.

The quartzose gravel and calcareous sand deposit on the summit is believed to be a western part of the Ogallala formation which, 40 miles to the east, constitutes the Llano Estacado of the Texas Panhandle. If this is correct and if the solution cavities and flowstone about the head of Nuevo Canyon are of the same age as the caves and surface exposures of flowstone elsewhere beneath the summit oldland and also if the Ogallala formation is Pliocene in age throughout, then it follows that the peneplain is either pre-Pliocene or had been made by the time the fill was complete, and that Carlsbad, New, and other caves are products of ground-water work in earlier stages of the peneplain cycle.

Fragmentary bones found in the vadose stream channel northeast of the Lunch Room in Carlsbad have been identified by C. L. Gazin of the United States National Museum as of *Nothrotherium*, a Pleistocene ground sloth. If they were carried into this part of the cave by a detoured surface stream, the channel and the deeper Pecos Valley, to which it must have discharged, are of Pleistocene age. This postulated deep stage of Pecos Valley is not to be confused with the original making of the lowland and the first appearance of the scarp. That original lowland extended on eastward into Texas beneath the great, high-plain Ogallala remnant—the Llano Estacado. The deep stage is a post-Ogallala valley or series of sinks with subterranean drainage, and it marks the reappearance of the scarp. The piedmont limestone gravel is post–deep stage.

A pre-Ogallala scarp must have been canyoned, and Ogallala sediment must have been deposited in such canyons. Relocation of modern canyons on old canyon sites appears to have destroyed the record of this. If it ever is found, the concept here advanced will become unassailable.

If the caves are of prescarp age and if the Ogallala formation later filled the eastern lowland and lapped over onto the reef summit, an early vadose experience of the caves must have been succeeded by a return to phreatic conditions. The spongework of some of New Cave's great stalagmites can be considered a record of this second phreatic experience. Failure to find a similar record in the other caves may mean only that their dripstone is younger or that their spongework in secondary limestone has been entirely covered by additions during the present vadose experience.

What were the conditions under which deposition of calcium carbonate ceased in some of the caves, was succeeded by deposition of calcium sulfate, and that followed by a return of lime deposition and a re-solution of the gypsum? A more fundamental question deals with the precipitation of the sulfate. Obviously, the familiar explanation of escape of CO_2 does not apply. For these questions the writer has no answers.

ACKNOWLEDGMENTS.—Closely associated with the writer in the field work and discussion on which this paper is based was Dr. Leland Horberg, of the University of Chicago. Thanks are gladly given him for suggestions, counsel, and a healthy skepticism until adequate evidence had accumulated. It is pleasant to record the wholehearted co-operation of the administrations of Carlsbad Caverns National Park and Lincoln National Forest during the field study. Particular thanks are due to Park Superintendent Donald Libbey, Park Naturalist Bennett Gale, Park Ranger Tex Worley, and Forest Ranger John McCollum, who volunteered as guides to and through some of the caves. The field expenses of the study were met by a grant from the R. D. Salisbury fund of the University of Chicago.

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GEOMORPHIC HISTORY OF THE CARLSBAD CAVERNS AREA, NEW MEXICO[1]

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ABSTRACT

The upland surface of the Guadalupe Range and three younger erosion surfaces in the Pecos lowland record progressively lowered base levels during the Pliocene and Pleistocene. Structurally, transverse consequent drainage in the range is attributed to superposition from a Pliocene (Ogallala) cover uplifted by block-faulting. Carlsbad and other caverns antedate the present topography and are believed to be related to a pre-Ogallala cycle of erosion.

INTRODUCTION

GENERAL STATEMENT

Carlsbad Caverns National Park in southeastern New Mexico is situated along the southeastern rim of the Guadalupe Range and overlooks the Pecos lowland to the east (fig. 2). The summit peneplain and high-level gravels, in the range, and terraces and valley fill, in the lowland, provide a basis for an interpretation of the main late Cenozoic events in the region. These events can be related to the time of cave development[2] and are of regional interest in their bearing on the genesis of the Pecos depression.

The area studied during 5 weeks of the summer of 1947 includes the central part of the Guadalupe Range and the adjoining Pecos lowland, east to the Pecos River (Carlsbad, West Carlsbad, Carlsbad Cavern West, Carlsbad Cavern East, and Malaga quadrangles of the U.S. Geological Survey) (fig. 3). Numerous traverses, which were made across the Pecos depression between Santa Rosa and the Texas boundary, provided a background of regional relations.

PHYSIOGRAPHIC SETTING

The physiographic boundary between the Pecos section of the Great Plains and the Sacramento section of the Basin and Range Provinces (Fenneman, 1930) follows the Reef escarpment at the southeast margin of the Guadalupe Range and essentially bisects the area studied (fig. 3). The range thus is the easternmost of the Basin ranges and is a fault block tilted to the northeast.[3]

Physiographic studies in the Roswell Basin and adjoining regions north of the area (Meinzer *et al.*, 1927, pp. 6–16; Fiedler and Nye, 1932, pp. 10–18, 25–41, 95–113; Morgan, 1938, pp. 9–17; Morgan and Sayre, 1942, p. 35) resulted in the recognition of five cyclical surfaces (fig. 1), from oldest to youngest: (1) the Sacramento plain, an upland surface west of the Pecos which truncates the structure on the north slope of the Sacramento Range and is correlated with the constructional surface of the High Plains to the east on the Ogallala formation; (2) the Diamond A plain, a subsummit rock plain, 400–1,300 feet below the Sacramento plain, which is correlated with gravel-capped mesas along the western edge of the Roswell Basin and with the Mescalero plain east of the Pecos; (3) the Blackdom terrace, a sloping

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[1] Manuscript received April 26, 1949.

[2] See companion paper by J H. Bretz.

[3] For a detailed treatment of the stratigraphy, structure, and geomorphology of the southern Guadalupe Range and adjoining areas see King (1948). This report was not available at the time that the present paper was written.

plain west of the Pecos underlain by limestone gravels and situated 55–90 feet above present drainage; (4) the Orchard Park terrace, a similar plain, 25–40 feet above present drainage; and (5) the Lakewood terrace, which is 20–30 feet above present streams and underlain by silty alluvium. All but the Sacramento plain are considered to be Pleistocene in age.

In the Carlsbad Caverns area it is possible to recognize the three lower terraces, and it is probable that the Guadalupe upland surface is equivalent to the Sacramento plain.

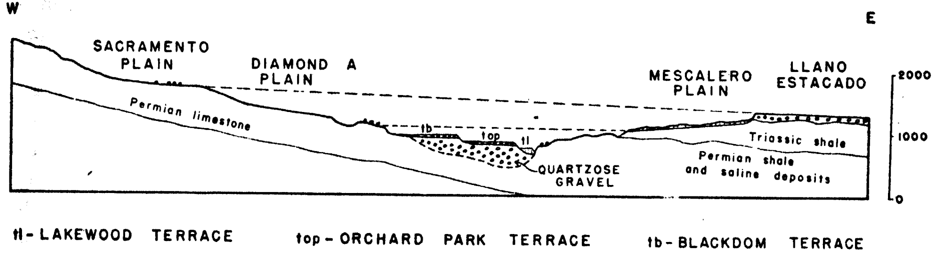

FIG. 1.—Erosion surface along Pecos lowland

GEOLOGICAL RELATIONS

The rocks of the Guadalupe Range are largely limestones belonging to the Permian Guadalupe series, which dip gently to the southeast and northeast. The mountain block consists of two marginal limestone uplands, which converge southward to give it a distinctive triangular shape. The northwest upland prong is underlain by limestones of both the Guadalupe and the underlying Leonard series, and the northeast prong ("La Barrera del Guadalupe" of Lang, 1937, p. 839) is underlain largely by Carlsbad and Capitan limestones of the Guadalupe series. The Carlsbad and Capitan limestones are stratigraphic equivalents and represent, respectively, back-reef and reef facies. Seven Rivers lowland, which separates the two uplands at the north, is eroded in an anhydrite wedge of the Guadalupe series (King, 1942, p. 553 and pl. 1).

The prominent Reef escarpment is a unique geomorphic feature. It is the exhumed front of a Permian reef. Southwest of Carlsbad Caverns the escarpment is due to massive southeast-dipping reef beds of Capitan limestone; northeast of the caverns it is produced by upper beds of Carlsbad limestone, which cover the reef and dip 10° southeast (King, 1942, p. 639).

The Pecos lowland is underlain by the Castile anhydrite, Salado halite, and Rustler limestone of the Ochoa series, which overlies the Guadalupe beds.

DRAINAGE AND CLIMATE

The entire area lies within the Pecos drainage basin and is drained by intermittent streams which enter the lowland from the southwest. The Pecos River and lower Black River are the only permanent streams.

The climate is semiarid, with a mean annual precipitation of 12.71 inches and a mean annual temperature of about 60° F. at Carlsbad. Most of the precipitation falls as showers during the summer months.

SUMMIT PLAIN IN THE GUADALUPE RANGE

DESCRIPTION

Almost the entire summit of the range is marked by accordant ridge crests and flat summit areas which descend to the northeast, truncating more than 3,000 feet of strata. In local areas truncation, although readily observed, is not a prominent feature (pl. 1, *A*). However, it is strikingly revealed by comparison of generalized contours of the upland with structure contours (fig. 2). Extensive flat summit areas occur on broad divides and on an upland plain covering several square miles in the vicinity of Queen. The reconstructed surface descends from elevations of about 8,000 feet near the southern apex of the range to elevations of about 4,000 feet near Carlsbad Caverns, an average slope of about 130 feet to the mile. This clearly is not an original slope and must be due to later tilting by faulting along the western margin of the mountain block.

The largest remnants of the plain have slight local relief, with shallow open valleys and broad low divides. The surface is cut off abruptly without gradational slopes at the rims of present canyons. Most of the surface is bare rock, thinly covered with weathering debris and alluvium. Solution-enlarged joints and sinkholes occur in many areas.

GRAVELS ON THE SUMMIT PLAIN

Siliceous gravel residuals occur in the upland plain in (1) the vicinity of Carlsbad Caverns; (2) near Jurnigan Draw, 7 miles northeast; and (3) on the uplands northeast and southwest of the mouth of Slaughter Canyon at elevations ranging from 3,750 to 5,950 feet. Pebbles of crystalline schist and foreign chert are reported[4] from Ares Peak, 22 miles west of Carlsbad Caverns at an elevation of 5,782 feet. In addition to these gravel residuals, conglomerate *in situ* is preserved in solution-enlarged joints at three localities near Slaughter Canyon—at the head of Nuevo Canyon, at a point just below the southern tip of the upland north of the mouth of Slaughter Canyon, and at the head of Yucca Canyon,[5] about 3 miles southwest.

More than 50 per cent of the conglomerate is made up of siliceous pebbles, which, in order of abundance, are chert, quartz, quartzite, and jasper. Most of the deposit is a granule gravel, predominantly of siliceous particles in a sandy, ferruginous matrix. Sandy and coarse conglomeratic facies were noted at the head of Nuevo Canyon.

Only a small fraction of the upland was examined for gravels, and it is probable that later detailed studies will find residuals to be widespread over much of the range.

The regional relations of the conglomerate and gravels to the quartzose conglomerate in the Pecos lowland (Meinzer *et al.*, 1927, pp. 9–10; Fiedler and Nye, 1932, pp. 35–38) and to Ogallala gravels of the High Plains indicate that a unit deposit filled an ancestral Pecos lowland and spread out over adjoining uplands as an extensive Tertiary cover over most of the Pecos section (Bretz and Horberg, 1949).

In the Carlsbad Caverns region this conclusion is supported by the occurrence of siliceous gravel residuals of similar lithology at all levels between exposures of quartzose conglomerate in the lowland and the summit of the range (fig. 3). Many occurrences doubtless represent re-worked Ogallala materials, but those at high elevations on the Guadalupe Range and the flanks of the Sacramento Range to the north approximate or lie above the projected High Plains surface; those at lower elevations occur on erosion surfaces, which indicate older gravels in the ranges as sources.

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[4] C. C. Branson, personal communication, 1947.

[5] B. T. Gale, personal communication, 1949.

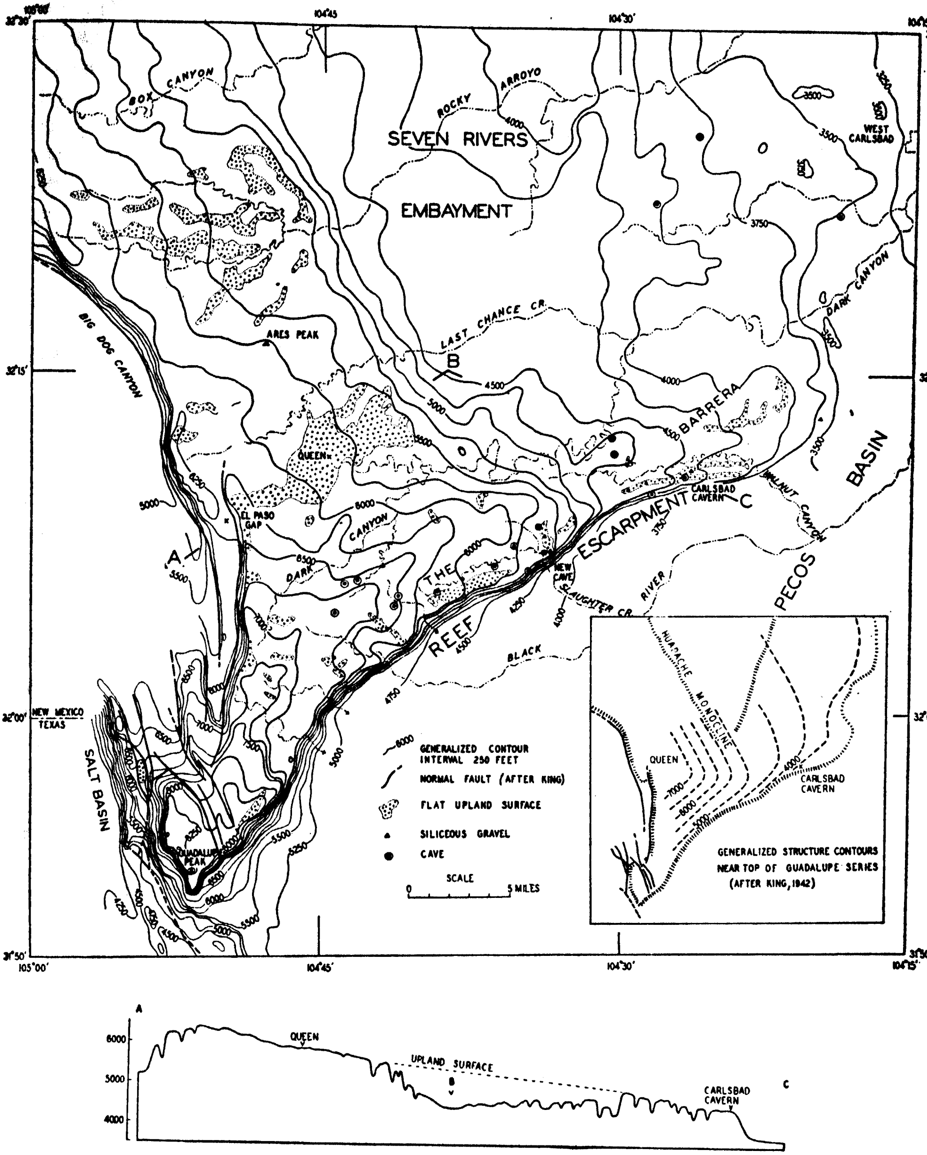

FIG. 2.—Generalized contour map of the Guadalupe Range

AGE AND CORRELATION

The gravels of the summit plain and their regional relations indicate that erosion of the surface probably was completed during the later stages of Ogallala deposition as the sheet of alluvium filled the lowland to the east and finally spread across the Guadalupe upland. If the plain had been cut earlier and dissected by streams on the pre-Ogallala surface, the critical evidence of pre-Ogallala valleys preserving valley fill has not been discovered. If the plain had been formed in post-Ogallala time, the transverse consequent drainage of the range would be difficult to explain, and also it is unlikely that the filled solution cavities and widespread gravel residuals would be preserved so extensively. The occurrence of remnants of Lower Cretaceous strata in the area suggest that important truncation of strata occurred in pre-Cretaceous time and that the present surface could have resulted from refaceting and lowering of an earlier extensive plain.

A correlation of the Guadalupe upland surface with the Sacramento plain in the Sacramento Mountains to the north is suggested by its comparable elevations above the Pecos lowland and similar gravel residuals (Bretz and Horberg, 1949). Uncertainties arise because of the absence of topographic maps for the region north of the Carlsbad area and the lack of knowledge of the tectonic history of the Sacramento Range.

A correlation of the Guadalupe surface with the Diamond A–Mescalero plain of the Roswell area is suggested by comparable levels of the Guadalupe upland north of Carlsbad and the Mescalero plain, 10 miles to the northeast (Oil City quadrangle). The projected levels lie about 175 feet above Pecos River. This approximates the 150-foot terrace in the area just to the north (Meinzer *et al.*, 1927, p. 6) which has been correlated with the Diamond A–Mescalero surface. It is believed, however, that the upland surface near Carlsbad has been reduced below original levels, as indicated by a break in slope north of Carlsbad Caverns (fig. 2). These lower uplands are not sufficiently uniform to warrant correlation with the Diamond A–Mescalero plain, and their accordance with that surface probably is coincidental. Correlation of the unreduced Guadalupe surface with the Mescalero plain would imply that the Mescalero is also an exhumed pre-Ogallala surface.

VALLEY FILL IN THE PECOS LOWLAND

QUARTZOSE CONGLOMERATE

The quartzose conglomerate of the Pecos depression is well known from exposures in the Roswell Basin (Meinzer *et al.*, 1927, pp. 35–38; Fiedler and Nye, 1932, pp. 35–38; Morgan, 1938, pp. 14–15), and the same deposit with similar lithology and structural relations is widely exposed in the Caverns area. Numerous exposures occur along the Pecos River, and more or less continuous ledges are present along the lower Black River for a distance of almost 30 miles (fig. 3).

The deposit is a firmly cemented conglomerate, composed largely of limestone pebbles but containing abundant foreign siliceous pebbles. The siliceous pebbles, in order of abundance, consist of quartz, chert, and quartzite. Granite, rhyolite, schist, basic igneous, and sandstone pebbles are minor constituents. Interbedded siltstone and sandstone were noted at a few localities. The exposures along Black River show increasing percentages of limestone and coarse material westward toward the mountain front. One of the best-exposed sections of the conglomerate in the area at Telltale

Bluff along Pecos River, 3 miles southeast of Carlsbad (fig. 3), is described in table 1.

Wherever the conglomerate is well exposed it shows evidence of disordered tilting, faulting, or downwarping with dips up to vertical (pl. 1, *B*). At two localities along Black River (N.E. ¼, S.W. ¼, sec. 12, T. 25 S., R. 25 E.; S.E. ¼, N.E. ¼, sec. 8, T. 24 S., R. 28 E.) local collapse by subsidence is indicated by centripetal dips. The Permian bedrock appears to be even more deformed, and this deformation in most places is disconformable rather than conformable with that of the overlying conglomerate.

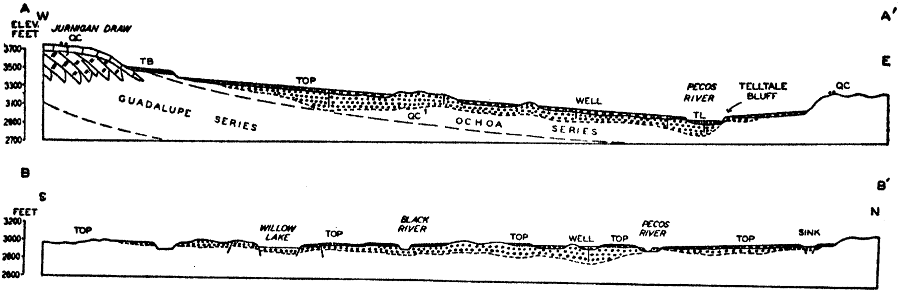

FIG. 4.—*A*, cross section along *AA′*, fig. 3; *B*, cross section along *BB′*, fig. 3

TABLE 1

QUARTZOSE CONGLOMERATE AT TELLTALE BLUFF ALONG PECOS RIVER
S.W. COR., SEC. 11, T. 22 S., R. 27 E.

Description of Rock	Thickness (Feet)	
	Unit	Cumulative
1. Caliche caprock, light-gray, dense, irregular layers, brecciated, concretionary; isolated siliceous pebbles and traces of limestone pebbles	3½	3½
2. Limestone-pebble conglomerate with chalky caliche cement and coatings; white; average diameter about 2½ inches; siliceous pebbles	8	11½
3. Limestone-pebble conglomerate, gray, dense, breaks across pebbles, irregularly bedded in channels and in low foresets; abundant siliceous pebbles; locally channeled into underlying silt	14	25½
4. Silt, red to brown, horizontally stratified; sandy, gravelly; pebble bands; abundant siliceous grains and pebbles; evidences tilting and minor faulting; rests with angular unconformity on underlying Permian limestone	26	51½
5. Limestone bedrock (Permian), dips steeply to southeast	....	....

The lack of structural continuity, the existence of recent sinks, and the presence of underlying evaporites point to solution-collapse as the cause of deformation.

Drillers' records of about fifty irrigation wells[6] in the area between Carlsbad and Malaga show that the conglomerate has a maximum thickness of over 200 feet. The materials penetrated include important thicknesses of sand, gravel, and conglomerate but reveal a higher proportion of silt than is indicated by surface exposures. The deepest part of the fill is continuous just to the west of

[6] Kindly provided by H. P. Burleigh, U.S. Soil Conservation Service, Carlsbad, N.M.

the present river, and there also appears to be an extension of deep fill westward along the present course of Cass Draw. The fill terminates south of Malaga, where bedrock is continuous across the floor of the present valley, and it is also separated by a bedrock barrier north of Carlsbad from a fill of somewhat greater thickness in the Roswell Basin. These relations indicate, as in the Roswell Basin (Morgan, 1938, pp. 11–12), that the depression in which the fill is now preserved is due largely to solution-subsidence rather than to erosion by the Pecos River. The subsurface data show that the base of the conglomerate was downdropped over 100 feet by subsidence. The continuity of the fill, however, suggests that most active solution may have occurred along the axis of former surface valleys.

There is no definitive evidence of the age of the quartzose conglomerate, as no fossils have been found. Previous writers (Meinzer *et al.*, 1927, p. 7; Fiedler and Nye, 1932, pp. 109–113; Morgan, 1938, pp. 14–15) tentatively considered it early Pleistocene. Its relations to high-level gravels in the area and along the Pecos lowland, north as far as Santa Rosa, however, indicate that it may represent a basal facies of the Pliocene Ogallala formation (Bretz and Horberg, 1949).

GRAVEL RESIDUALS IN THE PECOS LOWLAND

Siliceous gravel residuals above the level of exposures of quartzose conglomerate are widespread throughout the Pecos lowland (Bretz and Horberg, 1949). Within the area they occur in the mature uplands south of Black River at elevations ranging up to 4,700 feet (aneroid). The siliceous pebbles are similar in lithology to those in the quartzose conglomerate and the residuals in the Guadalupe Range. Limestone-pebble conglomerate with siliceous pebbles, found *in situ* at the northern end of the Yeso Hills (cen. sec. 12, T. 26 S., R. 24 E.) and at the north edge of upland (S.E. ¼, N.E. ¼, sec. 19, T. 24 S., R. 27 E.) (fig. 3) doubtless represents the parent-deposit from which the scattered siliceous pebbles were derived. At the latter locality the conglomerate occurs about 1 mile from, and 150 feet above, tilted ledges of quartzose conglomerate along the Black River. The similar lithology of the two conglomerates and the presence of residuals in the intervening area are strong evidence that a unit deposit once covered the entire area.

An isolated occurrence of debris composed of Washita (Lower Cretaceous) sandstone and fossiliferous limestone at the north end of the Yeso Hills (cen. N. ½, sec. 31, T. 25 S., R. 25 E.) is reported by Lang (1947, pp. 1472–1478). This material differs from the conglomerate, and associated siliceous pebbles are clearly differentiated from it. The material is believed to have been let down in a solution channel in the Permian fore-reef evaporites from once overlying basal Cretaceous beds. Cretaceous fossils in float in the Guadalupe Range and patches of sandstone near Jurnigan Draw also are reported by King (1942, p. 640).

PIEDMONT PLAINS AND TERRACES

BLACKDOM PLAIN

Remnants of the Blackdom plain occur as a frayed upland along the mountain front between Dark Canyon and Slaughter Canyon (fig. 3). The surface where it is best preserved slopes eastward at a gradient of about 50 feet per mile and is about 75 feet above Black River drainage. Northeast of Walnut Canyon the surface is bordered on the east by the Orchard Park plain and at many places is separated from it by a scarp 25–40 feet high. Southwest of Walnut Canyon the surface is extensively eroded, and its outer boundary is not so clearly defined.

The surface is underlain by limestone gravels which have a well-developed caliche caprock. Along the mountain front near Jurnigan Draw the gravels rest on a beveled bedrock surface. At the outer margin of the surface near the mouth of Walnut Canyon and Blue Spring, however, the areal relations indicate that the limestone gravels rest on a surface which truncates tilted quartzose conglomerate (fig. 3).

Scattered sinkholes and a fault in the limestone gravels near the mouth of Jurnigan Draw are evidence that some solution-subsidence followed formation of the plain.

ORCHARD PARK PLAIN

The Orchard Park plain is represented by a great, undissected, fan-shaped surface which has its apex in a re-entrant of the mountain front near the mouth of Dark Canyon, and by smaller terrace remnants along Pecos River, Black River, and Rocky Arroyo (fig. 3). The extensive surface heading in the Dark Canyon re-entrant has an eastward slope of about 45 feet per mile and near the city of Carlsbad is 30–40 feet above the Pecos River. Most of the irrigated land in the Carlsbad Basin is on this surface. The related terraces east of the Pecos are 30–50 feet above drainage, and those along Black River are 25–75 feet.

The surface of the plain is covered by brownish soil and alluvium, which in most places overlie calichified limestone gravel. The caliche zone does not appear to be so strongly developed as on the Blackdom gravels. Sandy and silty beds interbedded with the gravel were noted at a few localities. An exposure of the gravel along Pecos River at the east end of Wildcat Bluff (Carlsbad quadrangle) shows northeast-dipping foreset gravels, channel gravels, sand lenses, and interbedded brown silt and sand. Here and at several other localities south along the Pecos and along the Black River, the gravels overlie deformed ledges of quartzose conglomerate (pl. 1, *B*).

The thickness of the limestone gravel, as far as can be determined from exposures and well records, is not more than 50 feet and probably considerably less. This is also suggested by several islands of quartzose conglomerate and bedrock which rise above the level of the plain.

The uniform surface of the main plain is modified by shallow draws leading from canyons in the mountains; by Cass Draw, a broad depression on the plain; and by scattered shallow sinks. South of Black River the southward extension of the surface is interrupted almost completely by an area of recent solution-subsidence centering around Willow Lake (Malaga quadrangle). Beyond this area the surface continues south as a less uniform gravel-covered plain surrounding numerous residuals of older topography. To the west it merges with broad gradual slopes on bedrock in the adjoining uplands. Along the north flank of these uplands remnants of limestone gravels, derived from the Guadalupe Range to the west, were noted up to elevations of 100 feet above Black River (fig. 3).

GRADED PLAINS SOUTH OF BLACK RIVER

The graded plains in the mature uplands south of Black River merge with the Orchard Park plain on the north and east (fig. 3) and probably belong to the same cycle of erosion. The plains conform to present drainage lines and have concave profiles which break rather abruptly with clusters of residual hills, i.e., Cottonwood Hills and Yeso Hills (pl. 1, *C*). The slopes, in general, range from 75 to 100 feet per mile. In many places the surface clearly truncates bedrock structure, and it is probable also that locally the surface extends across fills in sinks and possibly across outliers of quartzose conglomerate. Youthful dissection is shown by the larger valleys and numerous arroyos slightly entrenched below the level of the plains. Recent sinks are not so conspicuous as they are in adjoining areas.

ORIGIN OF PIEDMONT PLAINS

The erosional, rather than constructional, origin of the Orchard Park plain is indicated by (1) the probable thinness of the gravel cover; (2) the beveled surface on deformed underlying bedrock and quartzose conglomerate; (3) the islands of quartzose conglomerate and bedrock surrounded by limestone gravels; and (4) the gradation of gravel-covered surfaces with gently sloping rock plains in the upland south of Black River. A similar origin for the Blackdom plain is suggested by characteristics 1 and 2. In the Roswell Basin the correlative surfaces were considered by Fiedler and Nye (1932, pp. 106–109) to be largely constructional features; but later Morgan (1938, p. 14), because of the thinness of the gravel cover, concluded that they were cut surfaces on older deposits.

The plains were graded to successively lowered base levels of the Pecos River and were formed on the weak Permian rocks and older valley fill of the lowland by streams from the mountains to the west. There was no retreat of the resistant Reef escarpment forming the mountain front; but, by the coalescence of cut plains and rock fans, the weaker sediments were effectively removed from the escarpment. The facts (1) that most extensive surfaces are related to the largest drainage lines in the mountains, (2) that the surfaces are restricted to soft rocks, and (3) that there has been no back-wasting of the mountain front indicates that lateral planation by streams was the dominant process.

The term "pediment," as now used (Bryon and McCann, 1936, pp. 148–152; Cotton, 1942, pp. 29–35), includes surfaces of this type, even though many "type" features assumed in some discussions are absent. The graded plains in the uplands south of Black River are restricted to drainage lines and might be termed "partial pediments." It is recognized that no clear distinctions can be drawn between plains of lateral planation, local peneplains on soft rocks, and the pediments as here described, so that a full discussion of the problem would be unwarranted in the present paper.

LAKEWOOD TERRACE

The Lakewood terrace is 15–30 feet above present drainage lines and is essentially continuous along the Pecos Valley (fig. 3). It also is recognized along Rocky Arroyo and with less certainty along upper Black River. Along smaller

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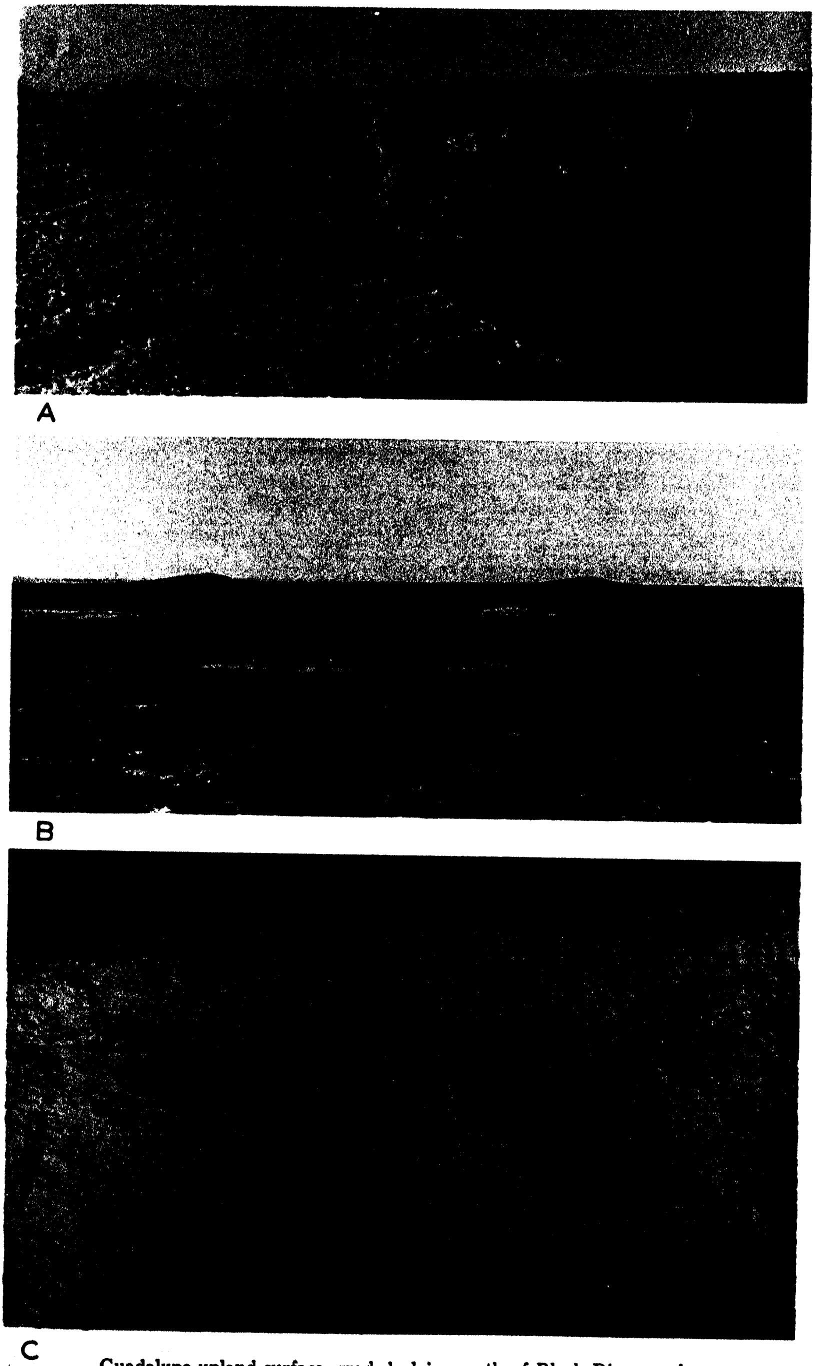
PLATE 1

A, Truncation of Permian limestones by the Guadalupe upland surface on south side of upper Black River Canyon. View south from Dark Lookout, Carlsbad Caverns West quadrangle.

B, Deformed quartzose conglomerate underlying undeformed limestone gravels along Black River, 3½ miles southeast of Carlsbad Caverns. Cut along U.S. Highway 62, Carlsbad Caverns East quadrangle.

C, Graded slopes and residual hills south of Black River, 8 miles west of Malaga (Malaga quadrangle). View south from Black River Valley.



Guadalupe upland surface, graded plains south of Black River, and deformed quartzose conglomerate.

tributaries it merges with recent alluvium. Slightly higher levels, up to 40 feet above drainage and 10–20 feet above the normal Lakewood, are represented by abandoned Pecos channels surrounding outliers of the Orchard Park plain.

The deposits underlying the terrace were noted at several localities along the Pecos and are composed almost exclusively of red-brown silt and silty sand.

PIEDMONT PLAINS SOUTH OF RATTLESNAKE CANYON

The piedmont plains fronting the mountains southwest of Rattlesnake Canyon (fig. 3) are more complex than those previously described, and their genesis and correlation are less certain. Superficially they appear to represent a series of piedmont alluvial fans in various stages of dissection. The underlying limestone gravels appear to be thick, and bedrock exposures of the fore-reef facies are limited to a few areas just in front of the Reef escarpment.

The plains, however, are not typical bajadas, in that the higher fan-shaped accumulations are in the interstream areas and are separated by youthful graded plains and valley strips which head at the mouths of major canyons. Two types of topography thus are present: the youthful plains along larger streams and the more dissected interstream areas. In places they are separated by definite scarps, and opposite Double Canyon an island remnant of the higher fan surface with a west-facing scarp is surrounded by lower valley strips. The lower surface is most extensive opposite Slaughter Canyon, where a broad sloping plain covers an area of several square miles. The average gradient of the lower surface is about 100 feet per mile and that of the higher, which is slightly steeper, about 110 feet per mile.

The lower surface merges with the present floodplain of Black River or is separated from it by a low scarp. This, together with the fact that it is slightly entrenched by present draws, suggests a correlation with the Lakewood terrace. The higher dissected plain, which is intermediate between the lower plain and remnants of the Blackdom surface to the north, could be considered equivalent to the Orchard Park plain.

The restriction of the lower plain to present major stream courses with the development of valley strips indicates origin by lateral stream planation, and the surface could be considered a partial pediment. The origin of the higher plain is less certain, as no evidence of beveling of older fill was noted. This evidence, however, could have been missed in the reconnaissance traverses made across the area and may be revealed by more detailed studies.

Numerous sinkholes, many with recent-appearing rims, are present on both surfaces.

CORRELATION AND AGE

The recognition of Blackdom, Orchard Park, and Lakewood surfaces in the area is based on relative elevations and their essential continuity along the Pecos and is believed to be strongly supported. Remnants of the higher Diamond A–Mescalero plain on the 150-foot terrace along the Pecos near the mouth of Rocky Arroyo (Meinzer *et al.*, 1927, p. 6) were not recognized in the area. The Mescalero plain, however, is extensive in the adjoining region east of the Pecos.

Tentative age assignments of surfaces in the Roswell area were made by Nye (Fiedler and Nye, 1932, p. 112) as follows: the Blackdom terrace to the Sangamon interglacial stage; the Orchard Park to the Peorian interglacial substage; and the Lakewood to a postglacial arid substage. Although the over-all Pleistocene age of the surfaces is evidenced by their regional relations to the Pliocene Ogallala formation, further correlations appear unwarranted. It is uncertain whether the surfaces were determined by diastrophic movements or by Pleistocene climatic changes; if they were climatically controlled, it is also uncertain whether erosion or aggradation occurred during the humid pluvials. Nye's concept of alternation of erosion and aggradation is based on the interpretation of the piedmont plains as constructional features. If, instead, these features are erosional, as proposed by Morgan (1938, p. 14) and in the present study, successively lowered, rather than fluctuating, base levels are indicated.

DRAINAGE EVOLUTION

GUADALUPE RANGE

The major northeast-flowing streams are consequent on the tilted surface of the fault block and in their lower courses are transverse across the Reef escarpment (Lang, 1937, pp. 896–898) (fig. 2).

Transverse drainage is especially notable in the cases of Rocky Arroyo and Last Chance Creek, which flow down the slope of the northwest prong of the limestone upland, cross the weak beds in the Seven Rivers lowland, and, instead of continuing north on the weak beds, cut through the east prong of the limestone upland to enter the Pecos.

Southwest of Slaughter Canyon the northeast consequent pattern gives way to a series of parallel short canyons, which drain southeastward normal to the Reef escarpment. Even in these canyons, however, the northeast trend is shown by tributary canyons, which almost exclusively drain northeast. Many of the canyons (i.e., Last Chance, North and South McKittrick, and Pine Spring) are beheaded by faults at the western scarp of the range. Dog Canyon drainage, at the base of the western scarp, follows a series of three tectonic depressions.

All the longer, northeast-trending canyons are characterized by incised meanders, largely of the ingrown type. Most striking examples are the Serpentine Bends of Dark Canyon (Carlsbad Caverns West quadrangle). The shorter, southeast-trending canyons in a few places have meander patterns of much smaller scale.

The accordance of major streams with fault-block structure and the tilted upland surface indicates that initial drainage formed during an early stage of block-faulting, or possibly along the axis of a monoclinal fold close to the present fault scarp. The transverse drainage is independent of secondary structures within the range and cannot be explained by headward erosion of subsequent streams or by antecedency. It is most logically explained by superposition, and the gravel residuals evidence an unconsolidated cover suitable for this process. The southeast drainage in the southern part of the range resulted largely from removal of weak beds fronting the Reef escarpment and in part from uplift along the Reef escarpment. The uplift is shown by abnormal dips along the escarpment (Lang, 1937, pp. 892–895; King, 1942, p. 587) and by southeast drainage trends in the Pecos lowland in the southern part of the area. Drainage of the southern tip of the range is directly southeast by way of Delaware River; farther north, southeast drainage appears to have been captured by headward erosion of Black River.

As the result of later block-faulting the initial consequent streams became entrenched and superimposed, incised meanders were formed, and Tertiary cover was stripped away except for scattered residuals.

PECOS LOWLAND

The Pecos depression is believed to have been produced by uplift of the ranges to the west, by solution-subsidence along a series of troughs, and to a minor degree by headward erosion (Lee, 1925; Lang, 1937, pp. 896–898; Morgan, 1942, pp. 27–35). Two stages, the present and a prequartzose conglomerate (Ogallala) stage, are indicated, and it is probable that the same factors apply to both. The major episodes of solution-subsidence shown by deformation of the bedrock formations and quartzose conglomerate occurred before and after deposition of the conglomerate. As a result of this solution and collapse, much of the soluble material underlying the lowland must have been removed because the overlying limestone gravels of the piedmont plains are undeformed except for minor recent sinks.

The present position of the Pecos within the area appears to be determined largely by later subsidence along the trend of the deepest fill of quartzose conglomerate. It may have been shifted slightly to the east of this fill by detritus which was shed from the mountains during erosion of the Orchard Park plain. As previously noted, upper Black River drainage probably is due to piracy of southeast-flowing consequents. The abundance of sinks in this area suggests that solution-subsidence was an important factor here also.

TIME OF CAVERN DEVELOPMENT

A study of seven caves in the Guadalupe Range clearly shows that they are older than the present topography, including both the present canyons and the much older upland surface (Bretz, 1949). This conclusion is based in large part on the following lines of evidence: (1) cavern passages transected by present canyon walls and upland surfaces, as revealed by present openings to the surface and more often by passages filled and choked off by debris; (2) caverns incommensurate in size with present drainage basins; (3) dripstone and flowstone deposits that are too large to be explained by the present thickness of overlying rock and drainage area; (4) dripstone and flowstone in surface rubble; and (5) features of the caverns indicating an early phreatic and later vadose stages in their history (Bretz, 1949).

The geomorphic relations thus indicate that the caves are among the oldest features in the area. Their intersection by the upland surface and the presence of gravel-filled solution cavities on summit areas indicate that they were formed either before or during an early stage of the Ogallala cycle. Their vertical range implies considerable relief in an earlier cycle, and the integration of subsurface drainage into the enormous passageways of Carlsbad Caverns indicates that the cycle reached a mature stage. These conditions appear to be fulfilled in large part by the Ogallala cycle of aggradation and an earlier cycle of erosion: the caves were initiated by phreatic circulation under a pre-Ogallala surface of important relief and became integrated into large passageways below the Guadalupe upland at a mature stage in the cycle; their upper galleries were transected as the upland surface was cut to lower levels toward the close of the cycle, and openings to the surface were filled with gravel as the rising Ogallala deposits covered the upland; the complex vadose cavern history of dripstone and flowstone deposits, roof collapse, and stream activity transpired as the Ogallala fill was removed from the lowland under the control of progressively lowered base levels of Pecos River.

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THE OGALLALA FORMATION WEST OF THE LLANO ESTACADO[1]

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ABSTRACT

Gravel remnants at various elevations over a wide area in southeastern New Mexico are related to a once extensive cover of High Plains Ogallala deposits. A correlation of the "quartzose conglomerate" fill along the Pecos depression with basal Ogallala is suggested.

INTRODUCTION

The Llano Estacado or Staked Plains of southeastern New Mexico and northwestern Texas constitute a clearly defined physiographic unit, rising above surrounding country by a marked scarp for almost its entire margin except the southern. This high, floorlike plain of 20,000 square miles is the but slightly modified surface of the Ogallala formation (Darton, 1898, pp. 732–742), generally regarded as largely of Pliocene age. Except for broad, shallow sinks, slightly entrenched stream courses, and some dune tracts, it is as featureless as when it was made.

The Ogallala formation of the Llano is not, however, of uniform thickness. A buried topography, chiefly in Triassic rocks, is expressed in thicknesses which range from a few tens of feet over pre-Ogallala hills and ridges to more than 400 feet in New Mexico and 550 feet in Texas over buried valleys.

Identified from Nebraska to Texas, the Ogallala formation, or group of formations, has been universally interpreted as a fluviatile apron deposited east of the Rocky Mountain front, the stream-transported waste containing minor members of eolian and lacustrine origin. The Llano Estacado portion, however, lies well to the southeast of the nearest southern Rocky Mountain range, the crystalline-cored Sangre de Cristos of Colorado and northern New Mexico. The mountain units of the Basin and Range Province, lying directly west of the Llano, are the Sierra Blanca, Jicarillos, and Capitan mountains of porphyritic lavas and the Sacramento and Guadalupe mountains of limestone. Most of the abundant quartzite, jasper, and vein quartz pebbles in the Ogallala of the Llano must have traveled southeasterly to reach their resting place. The general southeasterly slope of the Llano, averaging 10–15 feet to the mile, therefore appears to be the original depositional slope.

The western margin of the plateau-like plain is a scarp ranging up to 200 feet in height, slightly nicked in places by short westward-draining ravines but with its summit generally constituting a divide between the south-flowing Pecos on the west and the slightly entrenched easterly drainage of the plain's surface. In effect, this western margin is a low cuesta. Locally, buried hills of Triassic rocks appear in the retreating scarp, but for most of its length the Ogallala formation extends to the base.

The Pecos River, heading in the Rincon and Truchas ranges at the south end of the Sangre de Cristos, lies 12–40 miles west of the scarp and 300–1,000 feet lower. Several isolated areas of Ogallala,

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shown on the geologic map of New Mexico (Darton, 1928), lie west of the Pecos in the northern latitudes of the Llano (fig. 1). The largest one covers approximately ten townships. So far as the authors have seen, there are no scarped boundaries to these areas. South of these tracts, no Ogallala is mapped west of the Pecos for the 150 miles to New Mexico's southern boundary. It is generally believed, however, that the formation, before Pecos River erosion, extended west to these southern New Mexican ranges (Sellards *et al.*, 1943, p. 770).

This paper presents field evidence that the Ogallala formation once extended southward on the west side of the river at least to the Texas line, that Pecos drainage has removed it, save for a few small patches and widely distributed residual quartzose pebbles, and that the "quartzose conglomerate" in the bottom of the Pecos Valley is probably basal Ogallala instead of an early Pleistocene valley deposit.

The lithology of the Ogallala formation underlying the Llano plain has been described repeatedly as a complex series of gravels, sands, silts, and clays; the number of beds, the thicknesses, the sequences, and other characters varying so much from place to place that, in the lack of fossils contemporaneous with the deposition, even a generalized stratigraphic column has rarely been attempted.[2] The deposit is believed to have originated as a great alluvial plain or bajada on which various mountain-born streams debouched, deposited, wandered, divided into shifting distributaries, probably had temporary floodplain lakes, and probably provided the fines for associated wind transportation and deposition in the changing interstream areas. Some authors describe the basal portion as containing considerably more and coarser gravel than the remainder of the section; all recognize that a country of hills and valleys was invaded and eventually buried. Several recognize that the siliceous pebbles must have been contributed from mountains to the northwest and that the limestone pebbles came more largely from the west. Varying proportions of siliceous and calcareous constituents indicate shifting alternations of stream channels from the two contrasted source areas.

Induration of the Llano Ogallala is only partial. Calcite-cemented members are, by some, termed "caliche." That term, however, should be reserved for the most completely altered portion, the "caprock," which is interpreted by some as the uppermost sedimentary member of the formation, by others (including the authors) as an alteration product made in post-Ogallala time.

The surface of the Llano Estacado, the slopes of its bounding scarps, the deposits at scarp bases, and the exhumed lower country in front of them carry siliceous pebbles in places. A clear inference is that these pebbles are residuals and re-worked deposits from reduction of the Llano, surviving after disappearance of the matrix and associated calcareous pebbles of gravelly members, as the surface has been lowered by weathering and the scarps have been erosionally driven back from adjacent lowlands. In many places the siliceous pebbles are being released, by weathering, from the caliche caprock.

During nearly 2,000 miles of traverses on the Pecos lowland and Llano upland, ten crossings of the marginal scarps were made, seven of the western scarp, two of the eastern, and one of the northern. In all these crossings, gravel members or

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[2] See Sellards *et al.*, 1933, pp. 767–771, for summary.

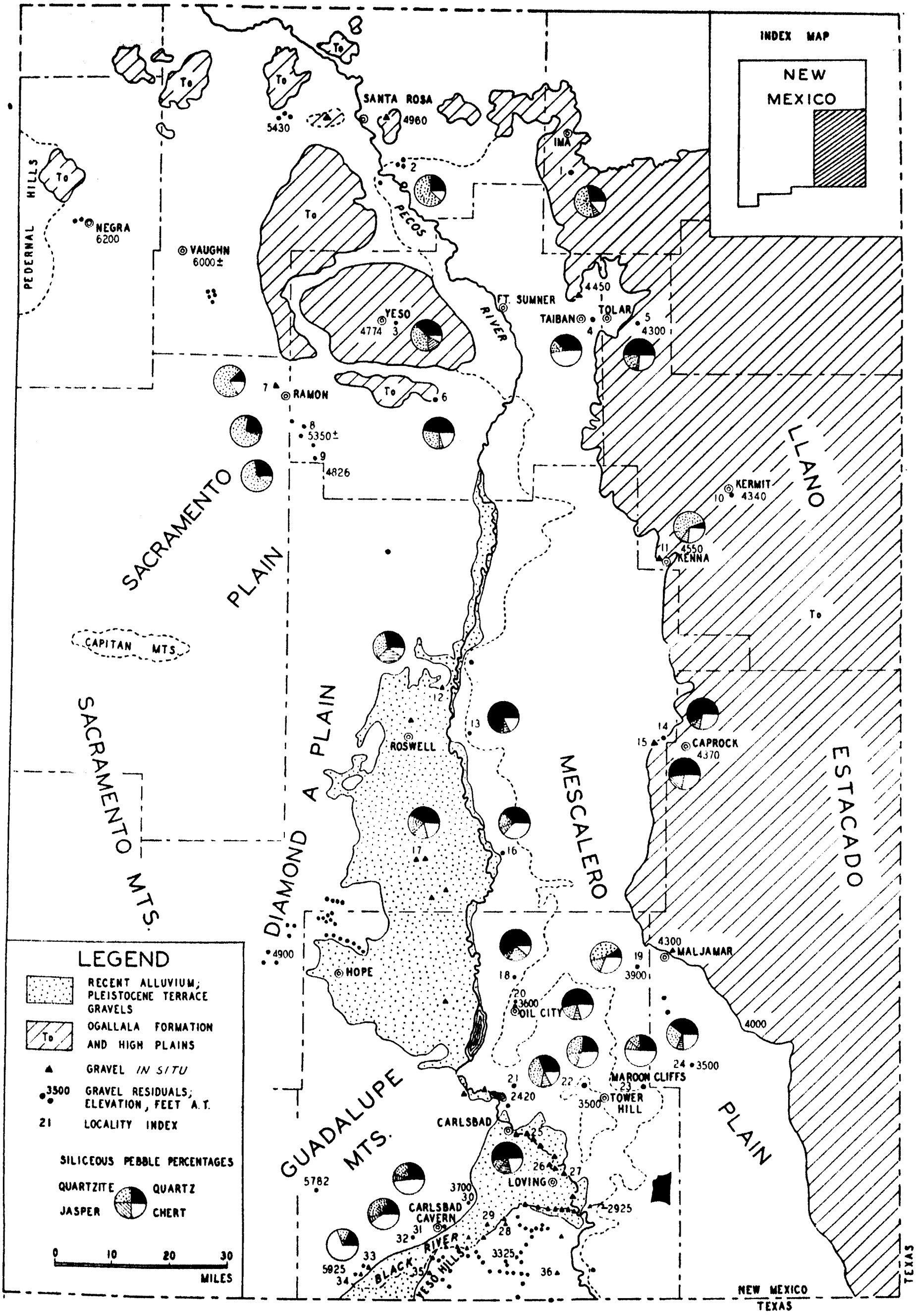

FIG. 1.—Ogallala formation and siliceous gravel residuals in southeastern New Mexico

slope debris showed pebbles of quartzite, vein quartz, jasper, granite, chert, porphyry, and lava, the siliceous pebbles being far more abundant than those of igneous rock, some strata being composed almost entirely of them. Carried by drainage from the western scarp and left behind by cliff retreat, these foreign pebbles are widely scattered over the Pecos lowland. The authors found 38 localities where they lie above all existing Pecos terraces but below and beyond the scarp bases (fig. 1). Most localities are on hill slopes and summits of the lowland's rough topography, and none was counted where modern drainage could have brought the pebbles from the scarp.

Because the regional slope, southward along the Pecos and southeastward across the Llano, ranges through 1,650 feet in the lowland and 2,000 feet in the upland, altitude figures for these pebble localities are meaningless unless map coordinates accompany them. The significant item is the relative altitude of a locality in the lowland above adjacent drainage ways and below the near-by scarp summit.

OGALLALA AND RELATED GRAVELS EAST OF PECOS RIVER

The lowland between Pecos River and the scarp, exposed by eastward retreat of the Llano margin, ranges up to 40 miles in width and for the most part is included in the Mescalero plain (fig. 1). Low cliffs and cuestas of tilted Triassic red sandstone in its relief are believed to have been inherited, in part, from the pre-Llano topography. More striking are sinks and saline lagunas formed since the exhumation. Red dune-sand deposits (Mescalero sands, Darton and Reeside, 1926, fig. 1, p. 413) also are of postexhumation origin. The linear draws and basins of this Triassic lowland parallel the strike and genetically are rows of sinks rather than stream-made valleys of the present cycle.

TOWER HILL

One of the most pronounced strike cliffs of the lowland is the Livingston "Ridge"–Maroon Cliffs–Nimenim "Ridge" scarp, lying about midway between the river and the Llano scarp and possessing more than 200 feet of relief between its summit and the sinks in Clayton Basin and Nash Draw, just west of it. During retreat of Maroon Cliffs, several isolated hills have been left behind, Tower Hill (Clayton Basin and Nash Draw quadrangles; fig. 1), 20 miles northeast of Carlsbad, being one. This hill's summit is 2 miles from the cliffs and 50 feet lower, but more than 100 feet above the intervening country. Thick, dense caliche caps the hill and covers at least the southern slope. Scattered through this caliche are good-sized pebbles of chert, jasper, and quartzite. Twenty-five miles to the northeast is the Llano scarp, its base less than 300 feet higher. Cuts along U.S. Highway 62, between Tower Hill and the Llano, show similar pebbles at intervals as far east as Laguna Gatuna. The easternmost exposure (locality 24, fig. 1) is of a gravel deposit *in situ*, most of its pebbles composed of limestone. The altitude here is no higher than the top of Maroon Cliffs, 500 feet below the Llano plain, 10 miles distant.

FADEAWAY RIDGE

This "ridge" (Oil City quadrangle; fig. 1) is an east-facing slope descending 200–250 feet to Burton Flat and having Clayton Basin still farther east, with its sink bottoms another hundred feet lower between it and the Nimenim "Ridge." Most of the Nimenim cliff brink is lower than the top of the Fadeaway slope. The

Llano scarp base, 30 miles to the east, is about 400 feet higher. The upland above and west of the Fadeaway slope is a little plateau 4–5 miles wide, the surface dropping off over Chalk Bluffs and McMillan Escarpment to the Pecos River, about 4 miles distant and 300–350 feet lower.

The surface of this plateau is sprinkled in places with siliceous pebbles, the altitudes ranging between 300 and 400 feet above the Pecos and the same distance below the scarp base, 25 miles to the east.

CAPROCK

The Mescalero plain east of Roswell (fig. 1), lying between the river and the Llano scarp, carries much eolian sand, and only one locality of siliceous pebbles was found, about 12 miles east of the river. The pebbles lie scattered on the surface and embedded in a caliche deposit about 375 feet above the Pecos and nearly 500 feet below the scarp base, 25 miles distant.

This lowland should have an abundance of such pebbles, for the cliff near the top of the scarp has the largest outcropping of siliceous pebbles that the authors saw in the western Llano scarp (locality 15, fig. 1). Gravel layers up to a foot thick in a brown silt are composed almost wholly of quartzose pebbles. An associated, firm, dense limestone layer also carries scattered siliceous pebbles, though no grouping into seams was found. This limestone has a curious nodular texture, the nodules tending to be of uniform sizes along any given stratum plane. They are thought to be traces of limestone pebbles, the original gravel at this level having had only a small percentage of foreign material.

Scattered, broken siliceous pebbles lie in the soil on the top of the scarp, released during disintegration of the caprock caliche. The authors hope to show, in another paper, that the caliche itself is a weathering product of earlier and different climatic conditions. If so, the scattered pebbles, before incorporation into caliche, came from vanished gravelly members lying somewhat higher than the present Llano surface.

KENNA AND KERMIT

One of the larger breakdowns of the Llano's western scarp is utilized by the Santa Fe Railroad and U.S. Highway 70 to cross the declivity between Kenna and Kenna Gap. It is wide and smoothly floored but has a very limited contributing drainage area. Sinks on top near the gap suggest that the breakdown may be largely of sink origin.

Members of the Ogallala formation are exposed in a highway cut near the bottom of the grade (locality 11, fig. 1), and a surficial caliche deposit is opened in a subsummit pit in the slopes of Kenna Mesa to the north.

Quartzose pebbles dominate over limestone in the gravel of the highway cut and are rare in the caliche exposed in the hillside pit. But, at Kermit station on the top of the Llano, 15 miles northeast of the Gap and 300 feet lower (down the dip slope), fairly large siliceous pebbles are embedded in the caliche of the plain's surface (locality 10, fig. 1). Like those at Caprock, they are taken to record the weathering-away of a dominantly limestone gravel whose calcareous content has made the caliche.

TOLAR

Tolar lies in the largest erosional interruption of the Llano's western scarp (fig. 1). More than 150 square miles of the summit plain's drainage has been pirated here from a southeastern course into Texas to become a southwest-draining tributary of the Pecos. The salient known as "Taiban Mesa," north of the gap, shows somewhat less than 100 feet of Ogallala overlying Triassic red beds in the cliff. A coarse conglomeratic member of the Ogallala, marked by channel unconformities, possesses limestone cobbles up to 12 inches in maximum dimensions. The nearest source appears to have been the Permian Chupadera formation, 60 miles to the west. The scarcity of quartzose pebbles also seems to indicate a dominant stream from the high country about Vaughn, west of the present Pecos Valley.

The Taiban Mesa summit is caliche-capped. Scattered siliceous pebbles in this caliche lie 400 feet or more above the bottom of the arroyo in the gap. They also occur in a white sand immediately beneath the caliche.

Ogallala sediments are similarly exposed in the arroyo bottom near Tolar, for the high and low points of the modern topography here coincide roughly with those buried by the Ogallala formation. Quartzose pebbles in stratiform arrangement show in the arroyo banks.

Gravelly waste, carried by the arroyo in flood, has been derived in large part from the eastern slopes of the mesa. In this waste, pebbles of igneous and metamorphic rock are common. An earlier alluvium overlies the lower Ogallala slopes, so closely resembling the Ogallala itself that the disconformity is almost essential for separating the two. Some strata in this alluvium contain rehandled siliceous pebbles.

OGALLALA AND RELATED GRAVELS WEST OF PECOS RIVER NORTH OF CARLSBAD

Thirteen separate areas covered by the Ogallala formation and lying west of the Llano are shown on the geological map of New Mexico (Darton, compiled, 1928), the westernmost of these areas lapping up onto the Pedernal Hills almost to the Pecos–Rio Grande divide (fig. 1). This tract reaches an altitude about 2,000 feet higher than the northwestern and highest part of the Llano, 80 miles to the east. Thus a gradient of only 25 feet to the mile would be necessary to connect this near-divide remnant with the Llano.

The authors have found the quartzose pebbles on the surface or embedded in caliche at five different localities on the four of these thirteen areas which they have traversed. In addition, they have found pebbles and gravel *in situ* in more than a dozen localities in the gaps among these Ogallala remnants, where the state geological map shows only Triassic or Permian rock. The highest are along U.S. Highway 60, west of Vaughn, 1,000–1,200 feet higher than the Llano and 75 miles distant by a route directly across the region's contour lines (fig. 1). The gradient of the connecting line in this case is about 15 feet to the mile.

The best showing of gravel *in situ* was found along the private road to K.E. Ranch, 2 miles northwest of Ramon (locality 7, fig. 1). A large pit has exposed 35 feet of well-stratified, sorted, and worn limestone gravel with a liberal admixture of quartzose pebbles. Compared with silicate and silica pebbles near Negra and Lucy, both close to the crystalline Pedernal Hills (fig. 1), these siliceous pebbles near Ramon indicate another source area and something more than a 50-mile trip from parent-rock. The gravel is sufficiently cemented to maintain vertical pit walls, but the induration is inadequate for ledge-making on the slopes where the deposit occurs. Fifty miles to the east along the regional slope and not much more than 200 feet lower is the Llano scarp summit. This descent is too low to accept as an original gradient of the Ogallala formation across the Pecos Valley. It can be increased to adequate proportions by considering the K.E. Ranch gravel as basal Ogallala and adding 500 feet to reach the original surface. But, because the average thickness of the formation in the Llano is reported to be from 200 to 300 feet and the depth of the K.E. Ranch deposit is not known, any picture of an adequate eastward-descending depositional slope seems improbable.

Limestone gravels with rare quartzose pebbles occur at lower elevations on dissected uplands at about the level of the Diamond A plain north and west of Hope (fig. 1). These gravels were believed by Fiedler and Nye (1933, pp. 38-39) to be younger than the "quartzose conglomerate" but were considered correlative by Morgan (1938, pp. 14-15). Their physiographic position and a thick caliche caprock indicate considerable antiquity, and it is not unreasonable to consider them as remnants of a local facies of the quartzose conglomerate-Ogallala fill.

OGALLALA AND RELATED GRAVELS IN THE CARLSBAD BASIN

The Pecos Valley from Roswell to the Texas line may be divided into two sections, the Roswell artesian basin to the north, floored with Quaternary valley gravels lying on the Chalk Bluffs and San Andres formations and structurally separated from a southern Carlsbad Basin section, which has an incomplete Quaternary gravel cover on the Rustler and Castile formations. Both sections have outcrops and well penetrations of the quartzose conglomerate, an indurated gravel, sand, and silt deposit beneath the younger and largely unindurated terrace gravels but above the Permian formations (pl. 1, *A*). This conglomerate has hitherto been considered an early Quaternary valley deposit in a post-Ogallala Pecos Valley (Meinzer, Renick, and Bryan, 1927, pp. 9-10; Fiedler and Nye, 1932, pp. 35-38; Morgan, 1938, pp. 14-15). Field evidence presented herewith inclines the authors to the interpretation of this conglomerate as basal Ogallala.

If the base of the quartzose conglomerate, penetrated by numerous wells in the Roswell (Morgan and Sayre, 1942, pp. 33-35) and Carlsbad basins,[3] is considered the base of the Ogallala and if it be assumed that no postconglomerate deformation has occurred, a former Ogallala fill of more than 1,300 feet along the Pecos depression is indicated.

In the Roswell Basin, earlier students have found the quartzose conglomerate cropping out in two localities at the north end, along two tributary streams from the west, and in the narrows at the south end where the Pecos crosses the Seven Rivers-McMillan escarpment (fig. 1). All outcrops are relatively low in the valley. Since the artesian basin has been so thoroughly studied, the authors made no attempt to find other outcrops of the conglomerate in it.

Lithologically, texturally, and in its sedimentary structures the quartzose conglomerate is identical with certain coarser phases of the Ogallala. Neither deposit, within the Pecos drainage, has yielded paleontological evidence of its age. Warned by the close similarity between original and re-worked Ogallala, the authors do not stress the resemblance as evidence of identity. The best evidence supporting their idea lies in the southern section and chiefly above the level of the valley terraces.

In this southern section is a triangular area of hilly land on Rustler and Castile rocks, south of the Black River and west

[3] Well records provided by H. P. Burleigh, U.S. Bureau of Reclamation, Carlsbad, 1947.

of the Pecos (fig. 1). Well-worn pebbles of durable siliceous rocks are widely scattered over it. They have been found most commonly on summits and higher slopes, where the originally associated, more readily weathered material has been removed. They range up to 950 feet higher than the junction of the Black and Pecos rivers, and 300 to perhaps 600 feet higher than the surface of the Llano plain to the east (fig. 1). This wide horizontal and vertical distribution has only one logical explanation: the entire triangular area has had a cover of Ogallala, and the contributing streams were not limited to drainage off the limestone Guadalupes to the west. That explanation becomes an almost inescapable conclusion when three hillside and hilltop remnants of conglomerate *in situ* are considered.

The highest part of the triangular area, the Yeso Hills, lies well south along the western edge, close to the Black River and some 200 feet higher. High on the northern slope are exposures of a limestone conglomerate, 150 feet above the river (locality 35, fig. 1; center sec. 12, T. 26 S., R. 24 E.; Carlsbad Caverns East quadrangle). It is far too high for terrace gravel, most of it is well indurated, and the exposures suggest that it is tilted. Its pebbles are well rounded and apparently record a longer trip than the 7 miles separating it from the Capitan reef scarp. Three quartzite pebbles were found in weathered waste of this conglomerate—a small number but difficult to explain except as derived from the indurated gravel.[4]

A low hill, 5½ miles directly west of the Pecos River and 200 feet higher, consists wholly of conglomerate (locality 36, fig. 1; corner of secs. 4, 5, 8, and 9, T. 26 S., R. 28 E.; Malaga quadrangle). The hill rises some 30 feet above the gently sloping plain surrounding it. Corroded limestone pebbles constitute 99 per cent of the surface waste, but the polished, brightly colored, siliceous pebbles are easily found. The surrounding plain is also sprinkled with them. The hill is nearly 300 feet lower than the Llano plain about 35 miles distant, but, being 100 feet higher than the highest river-terrace gravels, it appears satisfactorily explained as a remnant of a former Ogallala cover.

A mile and a quarter directly south of the Roy Forehand Ranch (locality 29, fig. 1; Malaga quadrangle) is an isolated hill composed mostly of Permian rock but possessing an abundance of siliceous pebbles and some of lava in its surface waste and ledges of limestone conglomerate with foreign pebbles on the summit. The hilltop is less than 1 mile from the Black River, but 150 feet higher. The river has cut a little gorge in the quartzose conglomerate at the ranch, 175 feet below the hilltop. The rock in the valley bottom and on the hilltop is so similar in every way that little imagination and

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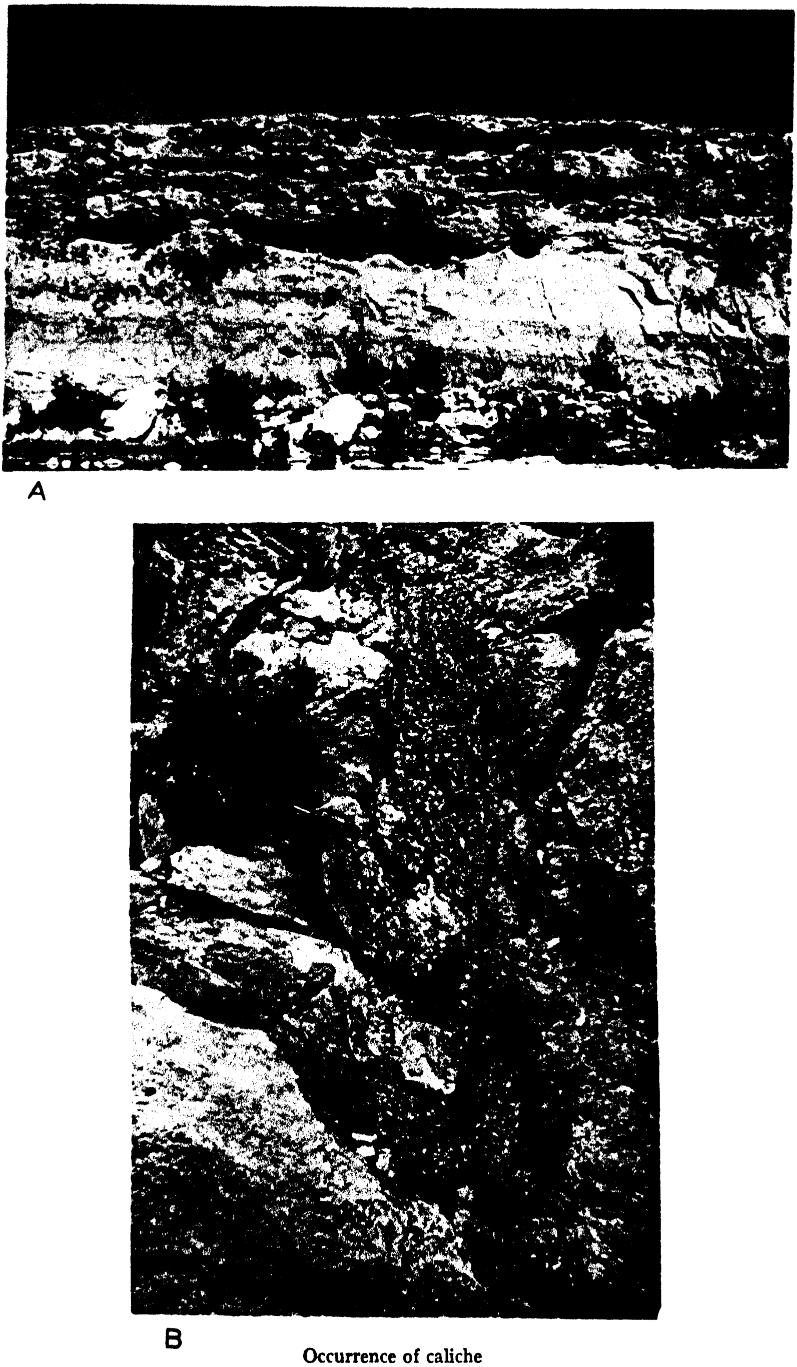
[4] The conglomerate differs lithologically from the Cretaceous sandstone and fossiliferous limestone remnants which Lang (1933) found about 3 miles to the north and interpreted as debris let down into a solution cavity in once overlying strata. Cretaceous fossils in float and patches of possible Cretaceous sandstone in place in the Guadalupe Range have also been reported by King(1942, p. 640).

PLATE 1

A, Angular unconformity between quartzose conglomerate and underlying Permian red beds along Pecos River, 4 miles north of Loving, N. M. (fig. 1; center S. ½, sec. 33, T. 22 S., R. 28 E.). Caliche caprock forms the ledge at the top of the conglomerate.

B, Quartzose conglomerate fills a solutionally enlarged joint in Capitan limestone near the crest of the Guadalupe Range north of Slaughter Canyon (locality 33, fig. 1; center N. ½, sec. 4, T. 25 S., R. 23 E.).



Occurrence of caliche

still less daring are required to see in this locality a record of a once-continuous indurated limestone conglomerate from channel floor to summit ledges (fig. 2), a former river gravel supplied largely from the western uplands but containing contributions from the Rocky Mountains, more than 200 miles to the north.

Were there only the hilltop conglomerate ledges, one who accepts preceding interpretations would be inclined to consider it a remnant of basal Ogallala. Were there only the river exposures, one familiar with the quartzose conglomerate along Black and Pecos rivers would unhesitatingly identify it as that deposit. The authors are strongly inclined to think that the two exposures are but parts of the same formation.

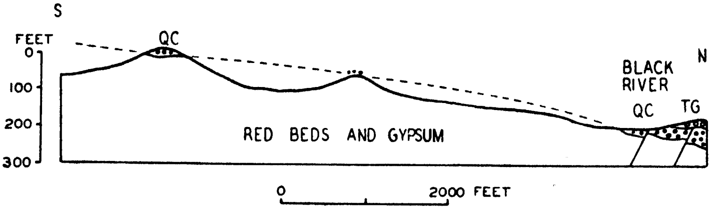


FIG. 2.—Relations of siliceous gravel residuals to the quartzose conglomerate near Forehand Ranch (Maluga quadrangle, New Mexico). *QC*, quartzose conglomerate and siliceous residuals; *TG*, terrace gravel.

Blue Spring (locality 29, fig. 1; Carlsbad Caverns East quadrangle), although below terrace levels of Black River Valley, is a large ground-water discharge from Permian limestone and/or quartzose conglomerate, both of which crop out in the low hills constituting the spring alcove. The conglomerate here, $1\frac{1}{2}$ miles from the river, lies more than 100 feet above it and has the same altitude as that on top of the Forehand hill, less than 5 miles away.

The foreign pebbles of both Forehand outcrops and of Blue Spring ledges came down along the Pecos from the north. How did they get 15 miles *up* the Black River Valley? The explanation appealing most strongly to the authors is that the quartzose conglomerate was deposited before a Black River Valley existed.

From Herradura Bend southward, the Pecos River loses its broad lowland floodplain and, thence to Texas, traverses several narrow stretches with quartzose conglomerate and Permian red sandstone bluffs. One bluff at Herradura Bend (locality 26, fig. 1; Carlsbad quadrangle) is 60 feet high and composed entirely of the conglomerate formation, as is also the channel floor. The opposite bluff is only half as high because the river for a short distance here flows along the slope of a former hill of the conglomerate that later became buried in younger terrace gravels from which the Pecos has been superimposed. The upper 10–30 feet of the hill still stands above those younger gravels. A clean section at the crossing of U.S. Highway 31, $\frac{1}{2}$ mile below Harroun Dam, shows Permian red beds in the river channel, overlain by 10 feet of red sandstone containing quartzose pebbles and that by 20 feet of gray lime-cemented conglomerate, also with abundant foreign pebbles (locality 27, fig. 1).

This is as low as any exposure of the quartzose conglomerate seen during this study, more than 400 feet lower than the outcrops at Blue Spring, 16 miles distant up Black River Valley. Pecos and Black River courses between the two localities afford many exposures, spaced close enough to carry conviction that the formation is continuous. To bring the foreign pebbles to Blue Spring, the region must have been a uniform plain of fluviatile deposition, in the aggradation of which Rocky Mountain water participated. The constrictions imposed by present valley ways and the direction of flow determined by present gradients did not exist when the quartzose conglomerate was deposited. The flat top of the hill south of Herradura Bend is not a depositional surface. Assuming that no postconglomerate local warping or downfaulting has occurred, this flat hilltop is 350 feet below the level required to carry the siliceous pebbles to the deposit at Blue Spring. To reach up to the projected level of the Llano plain across this locality, the former Ogallala fill here must have been 600 feet thick.

OGALLALA RESIDUALS IN THE GUADALUPE RANGE

In five different places, worn quartzose pebbles have been found on the top of the pronounced Capitan Reef scarp, by which the mountains break off to the Black River lowland (fig. 1). A few were picked up on the summit just south of Jurnigan Canyon (locality 30, fig. 1; Carlsbad quadrangle), a large number lie from $\frac{1}{4}$ to $\frac{1}{2}$ mile east of the entrance to Carlsbad Caverns (locality 31, fig. 1; Carlsbad Caverns East quadrangle), and a few were found on the upland, 3 miles southwest of the cavern entrance. At all these places the pebbles lie essentially on bare rock surfaces and have no associated limestone pebbles.

The other reef-summit localities are on either side of the deep notch of Slaughter Canyon in the scarp near the heads of two minor canyons, Nuevo (locality 33, fig. 1) and Yucca[5] (locality 34, fig. 1; Carlsbad Caverns West quadrangle). At each, the quartzose pebbles occur in remnants of an indurated gravel and sandstone filling in opened joints and solution cavities of the Carlsbad and Capitan limestones (pl. 1, *B*), as well as scattered over the bare surface or throughout the thin soil. The display of thousands of these red and yellow polished pebbles on the mesa summit at the head of Nuevo Canyon is spectacular. More than 50 per cent of all pebbles in the sandstone and conglomerate here are siliceous.

Possible Ogallala residuals, consisting of foreign pebbles of schist and chert, have been reported[6] from Ares Peak (Texas Hill quadrangle) at an elevation of 5,782 feet A.T., in the central part of the Guadalupe Range. This is 16 miles north-northwest of the Nuevo Canyon locality and about 260 feet higher. It is 240 feet lower than the pebbles at the head of Yucca Canyon.

The pebbles of the Jurnigan Canyon locality lie 750 feet above the Harroun Dam exposure of quartzose conglomerate in almost the same latitude. The residual pebbles east of the cavern entrance lie about 1,100 feet higher than the conglomerate at Blue Spring, almost directly east. Those lying 3 miles southwest of the cavern are about 1,400 feet higher than the outcrop at the spring. The pebbles and conglomerate *in situ* at the Nuevo Canyon locality are 1,600 feet higher than those on the Yeso Hills, only 8 miles distant to the southeast and the highest found in the before-mentioned triangular area. At the Yucca Canyon locality they are 2,000 feet higher than in the Yeso Hills.

These siliceous pebbles on the summit

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[5] The authors did not see this locality. It was discovered and reported to them by Bennett Gale, park naturalist, after the close of the field season.

[6] Carl Branson, personal communication.

of the reef scarp are considered to be Ogallala residuals and certainly are Rocky Mountain contributions. There was, by this view, a continuous southward and southwestward downgrade to the Ogallala alluvial plain along the foot of the Guadalupes, and the concept of an Ogallala bajada sloping eastward from them to the Llano (Sellards, 1933, p. 770) is incorrect. The Capitan Reef scarp is believed to be a product of pre-Ogallala erosion and to have been buried in the mounting waste toward the close of this Pliocene aggradation.

The scarp at the Yucca Canyon locality is nearly 1,500 feet high, and the younger gravel at its base slopes down another 700 feet to the approximate level of quartzose conglomerate outcrops near by in Black River Valley. Either the Ogallala formation in front of the scarp was 2,200 feet thick, or subsequent solution-collapse or warping has occurred to alter original levels. Warping along the line of the scarp seems supported by a comparison of altitudes of Ogallala remnants and residuals north of Roswell and west of the Pecos with the Yucca Canyon residuals and remnants. From Vaughn southward to Yucca by valley routes is about 200 miles, yet the descent, as read from the lingering quartzose pebbles, is only 100 feet. A gradient of 12 feet to the mile is secured by considering the Vaughn gravel as basal and equivalent to that along Black River, 2,200 feet below the Yucca remnants. But excessive thicknesses and improbably low gradients may be avoided by accepting some post-Ogallala uplift of the reef scarp relative to the adjacent lowland. This could be related to known Pleistocene movements of the Guadalupe block which faulted older alluvium along the west front of the range (King and Fountain, 1944). Differential uplift in this part of the scarp also seems required if the pronounced flatness of the top of the reef here is referred to the summit peneplain. The northwestward slope of this flat between Yucca and Nuevo localities is 100 feet to the mile.

Nevertheless, if the quartzose conglomerate is Ogallala, the thickest part of this formation lay along the present course of Pecos Valley, and this requires the concept of a pre-Ogallala Pecos Valley. The idea is not difficult to accept, for the valley lies along the strike and over the edges of the highly soluble evaporites of the Rustler and Castile formations. Sinks are actively growing today along this belt or are so recent that waste from their steep cliffs has not yet filled them. Also, most exposures of valley-bottom quartzose conglomerate show fracturing and strong tilting of big blocks because of solution-subsidence. Lee's idea (1925) that the course of the modern Pecos has been established by coalescence of sinks along the belt of upper Permian evaporites needs to be used also in interpreting the pre-Ogallala topography.

The induration of the quartzose conglomerate equals anything seen in the Llano gravels and far exceeds that of the overlying valley-terrace gravels. Generally, the matrix is as firm as the included limestone fragments, and the pebbles are impossible to crack out entire. Even river abrasion produces a smooth transection across matrix and pebbles. This cementation is not alone a matter of age; it implies also an original deeper burial than the terrace gravels have ever provided.

PEBBLE COUNTS

Pebble counts of the unweathered gravels (table 1) indicate the variable composition of the deposits. The gravels from the Ogallala formation and possible

Ogallala remnants show a dominance of limestone (38.6–60.3 per cent) compared with the quartzose conglomerate (2.0–11.5 per cent), which is composed largely of siliceous pebbles. However, all the quartzose conglomerate samples are from the axial portion of the Pecos Valley, and there are much higher limestone percentages in its remnants along the western margin of the basin. Exposures of the conglomerate along Black River, which are continuous with those along the Pecos, show overwhelming dominance of limestones. It seems just as probable that the concentration of siliceous material was brought about by an early valley phase of Ogallala deposition as by later Pleistocene re-working of Ogallala gravels.

Siliceous pebble percentages from Ogallala and related gravels and from the quartzose conglomerate are given in table 2 and shown graphically in figure 1. Again wide variability is indicated, but with close similarities in adjoining deposits, i.e., localities 7, 8, and 9 near Ramon and localities 31, 32, and 33 near Carlsbad Caverns (fig. 1). Three widely separated samples from quartzose conglomerate along the Pecos Valley axis (localities 12, 17, and 25) are strikingly comparable. The gravels from the Llano and the Mescalero plain differ from place to place. In general, there appear to be higher percentages of chert and jasper in all the deposits in the southern half of the region, probably because of proximity to Permian and older limestones in the Sacramento and Guadalupe mountains to the west.

TABLE 1

TOTAL PEBBLE PERCENTAGES OF UNWEATHERED OGALLALA AND OGALLALA (?) GRAVELS AND QUARTZOSE CONGLOMERATE

TYPE OF PEBBLE	LOCALITY (FIG. 1)							
	Ogallala and Ogallala (?)					Quartzose Conglomerate		
	1	2	4	7	11	12	17	25
Quartz	6.4	6.5	18.3	5.1	2.4	28.0	37.5	45.4
Jasper	1.7	....	....	....	3.4	28.7	15.9	12.6
Quartzite	11.5	16.9	6.8	29.1	20.3	8.0	18.2	11.5
Chert	3.8	3.4	23.5	6.1	9.1	29.3	19.3	18.6
Granitic	14.1	22.8	7.8	6.8	10.1	3.3	6.8	0.5
Basic	0.8	0.7	....	2.8	13.0	0.7	....	....
Limestone	60.3	46.2	42.9	43.0	38.6	2.0	2.3	11.5
Sandstone	1.2	3.4	0.5	7.1	2.9	....	....	....

CONCLUSIONS

1. The relations of gravel remnants in the region to the Ogallala formation indicate that the Ogallala cover once extended south at least to the Texas line and covered all but the highest portions of the Sacramento and Guadalupe ranges. This conclusion is based not on attempted correlation of all gravels in the region as Ogallala, because many, as on the Mescalero plain, are clearly re-worked, but on their occurrence in physiographic positions and at elevations which preclude re-working from an eastward-retreating Llano scarp. Remnants on the Sacramento plain and the

Guadalupe upland are at elevations accordant with or above the High Plains surface; those on the Diamond A plain and the upland south of Black River are on surfaces which slope eastward to the Pecos and, if re-worked, must have been derived from Ogallala deposits once present to the west rather than to the north or east.

TABLE 2

SILICEOUS PEBBLE PERCENTAGES FROM OGALLALA AND RELATED GRAVELS AND FROM QUARTZOSE CONGLOMERATE

Locality (Fig. 1)	Quartz	Quartzite	Jasper	Chert
1*	27.3	49.1	7.3	16.4
2*	24.4	62.8	....	12.8
3	36.9	39.4	16.9	6.9
4	37.7	14.0	....	48.4
5	49.1	21.8	3.6	25.5
6	48.3	26.3	6.3	19.1
7	12.5	72.5	....	15.0
8	24.5	72.2	2.2	1.1
9	28.1	66.3	....	5.6
11	6.9	57.5	9.6	26.0
12	29.8	30.5	31.2	8.5
13	72.2	2.2	7.2	18.3
14	60.7	0.5	8.2	30.6
15	54.2	....	18.1	27.7
16	38.4	15.2	8.0	38.4
17	41.2	20.0	17.5	21.3
18	61.1	4.7	22.6	11.6
19	9.4	45.3	15.1	30.2
20A	54.2	18.4	6.8	20.5
20B	54.7	11.0	12.0	22.3
21	29.5	40.0	13.3	17.1
22	24.3	27.0	16.2	32.4
23	26.5	12.2	10.2	51.0
24	39.2	29.4	6.9	24.5
31	25.6	17.1	8.1	49.2
32	18.8	34.4	6.2	40.6
33	17.6	13.5	0.7	68.2

* Unweathered gravels with many crystallines.

2. The relations of the gravels to the quartzose conglomerate in the Carlsbad area, especially in the Black River drainage basin, indicate that a single deposit buried a land surface which was much like the present and had a relief of almost 3,000 feet. The occurrence of gravel remnants of similar lithology at all elevations between Pecos River levels and summits in the Guadalupe Range is the strongest evidence for this view.

3. It follows from the preceding conclusions that an ancestral Pecos depression, probably formed jointly by warping and solution-collapse, existed in pre-Ogallala time and that the present depression, enlarged by retreat of the Llano scarp and modified by later deformation and solution-subsidence, is in large part a resurrected feature. It also follows that the "Reef escarpment," which forms the southeastern margin of the Guadalupe Range, has been exhumed at least twice: first by stripping of weak Permian fore-reef deposits and later by removal of the Ogallala fill.

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CALICHE IN SOUTHEASTERN NEW MEXICO[1]

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ABSTRACT

Caliches with similar zonal profiles lie on the Pliocene Ogallala formation of the Llano Estacado and on three younger surfaces along the Pecos depression. Physiographic relations, the caliche profile, and other features indicate that the caliches were developed by soil-forming processes from parent-materials consisting largely of limestone gravels. The caliches and associated solution features provide a partial record of climatic changes since middle Pliocene time.

INTRODUCTION

TERMINOLOGY

Throughout the Southwest and the southern High Plains the term "caliche" is applied to calcareous caprocks, soil hardpans, and earthy or porous materials which occur at the surface or at shallow depths below the soil. Impurities are present in variable amounts, and siliceous and ferruginous materials may be associated. In other areas the term also has been used for aluminous, siliceous, ferruginous, and saline deposits; for sediments cemented and impregnated with calcium carbonate, i.e., mortar beds; and for incipient lime accumulations in the subsoil. This broad usage endangers the usefulness of the term, and it is desirable, if possible, to restrict the term to materials of calcareous composition (Sayre, 1937, p. 66), as indicated by the derivation of the word, and specifically to those calcareous materials which are formed in the zone of weathering. Subaqueous deposits, spring deposits, and mortar beds thus would be excluded.

The term "caprock" is used in this region to describe the uppermost zone of the caliche profile. It is a dense limestone only a few feet thick in most places, but topographically prominent because of greater resistance to erosion than the deeper zones have. Although it is a "layer" and may have bandings and horizontal partings that suggest minor strata, its outstanding structure is that of a self-breccia, recemented by massive sandy caliche and by additions of calcareous laminae which wrap around the fragments and occupy interstitial space of larger volume than that of the fragments. Thus the fragments commonly are completely isolated in the matrix. In the type of caprock herein denominated "Rock House," the brecciation is compounded to record two or more cycles of fragmentation and rehealing (pl. 1, *A*).

NATURE OF STUDY

The present study is based primarily on field observations made during a series of traverses across the Pecos depression and surrounding highlands over an area extending from the Texas boundary north to Santa Rosa, New Mexico (fig. 1). Additional observations were made of caliche in western Texas and southern Kansas. Selected specimens were examined in the laboratory, but a comprehensive study of them was not attempted.

The area is of particular interest in testing hypotheses for the origin of caliche because that material is extensively distributed on four different physio-

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[1] Manuscript received May 6, 1949.

graphic surfaces and is derived from similar parent-materials, largely limestone gravels. The present paper describes the deposits and attempts to interpret their origin.

CLIMATE OF THE AREA

The area is semiarid with a mean annual temperature of about 52° F. and an average precipitation of 12–18 inches. Rainfall is greatest during July, August, and September and often occurs as heavy downpours with local thunderstorms. Evaporation is rapid as a result of low humidity and frequent high winds (Kincer, 1941).

GEOLOGIC AND PHYSIOGRAPHIC RELATIONS[2]

The Pecos Valley is a broad north-south depression bordered on the west by the Guadalupe and Sacramento mountains and on the east by the Llano Estacado of the High Plains. The mountains are composed largely of Permian and Pennsylvanian limestones, which dip gently eastward. Permian evaporites and Triassic red beds underlie Pleistocene and Recent deposits along the depression, and the Llano upland is underlain by sands and gravels of the Pliocene Ogallala formation (Darton, 1928*a*).

The High Plains of the Llano are believed to represent a constructional surface correlative with the Sacramento erosional plain in the Sacramento Range (Fiedler and Nye, 1932, p. 97). Lower erosion surfaces within the Pecos depression include, in descending order: the Diamond A–Mescalero plain, Blackdom plain, Orchard Park plain, and Lakewood terrace (fig. 1). Well-developed

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[2] See accompanying papers by the same authors in this issue of the *Journal* for further details.

PLATE 1

A, Brecciated caprock composed of dense caliche fragments and siliceous pebbles (*A*), in a sandy caliche matrix ("Rock House" structure). The fragments show solutional modification and have a thin rim of manganese oxide. From caprock near Rock House, N.M. (fig. 1; Malaga quadrangle; NE. ¼, sec. 8, T. 26 S., R. 29 E.). Specimen is 8 inches long.

B, Concretionary mass of dense banded limestone in chalky caliche zone above Ogallala deposits. Western margin of the Llano Estacado at Maljamar, N.M., along N.M. Highway 83.

C, Banded travertine from concretionary mass similar to the above. Llano scarp along N.M. Highway 31, 12 miles north of Maljamar, N.M. Specimen is 3 inches long.

PLATE 2

A, Caliche zones: (*1*) dense, banded caprock above hammer; (*2*) brecciated caprock, prominent ledge below hammer; (*3–4*), chalky caliche with cupped pebbles below ledge; (*5*) limestone pebble gravel, the parent-material, at the base of cut (see fig. 3). Blackdom plain, 1 mile northeast of Hope, N.M. (fig. 1).

B, Cupped-pebble zone at above locality.

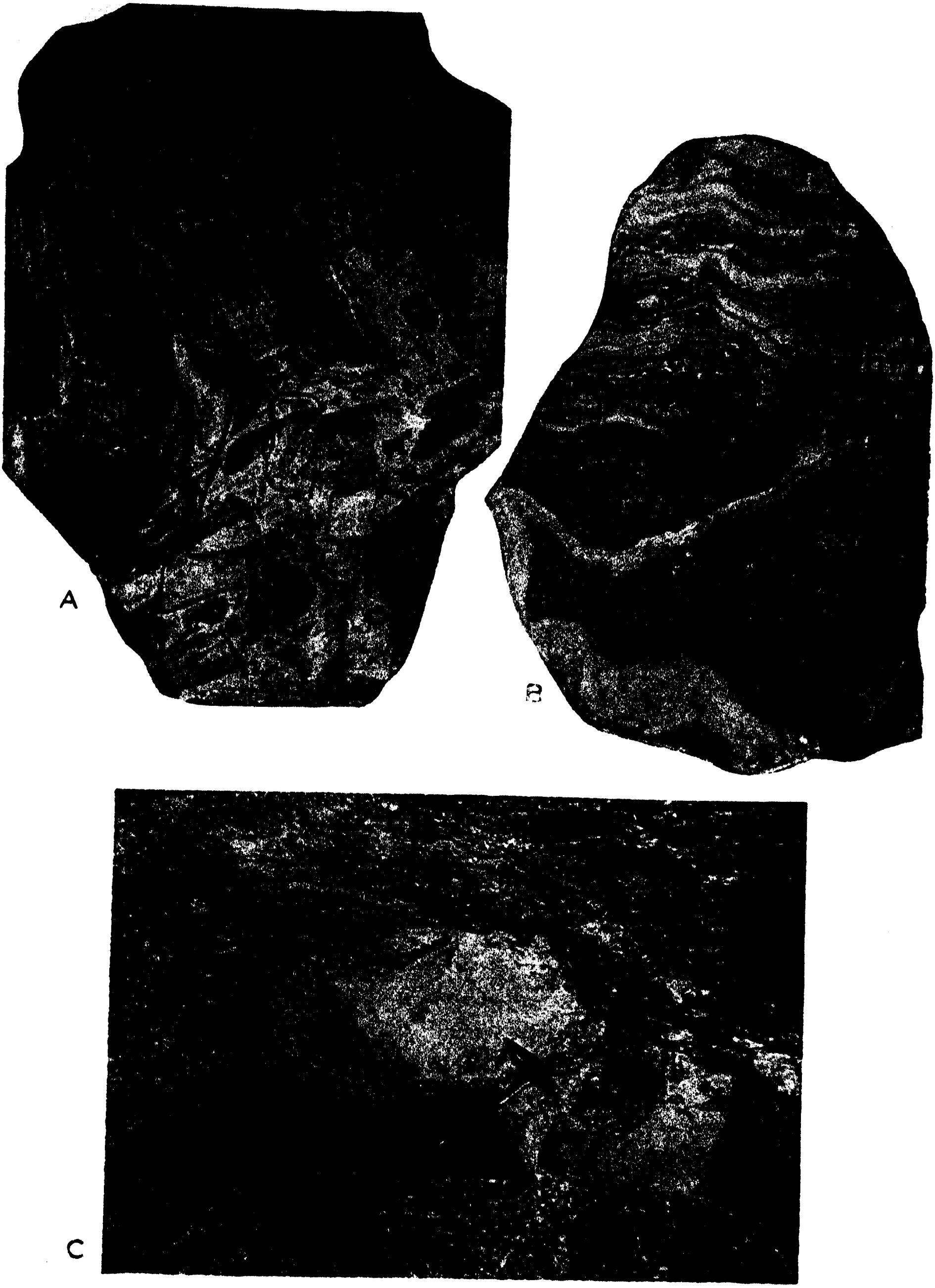
PLATE 3

A, cross section showing caliche crust on cupped limestone pebble; *B*, *C*, and *D*, upper surface of cupped limestone pebbles. From cupped-pebble zone in caliche pit, 1 mile northeast of Hope, N.M. The largest specimen is 3 inches long.

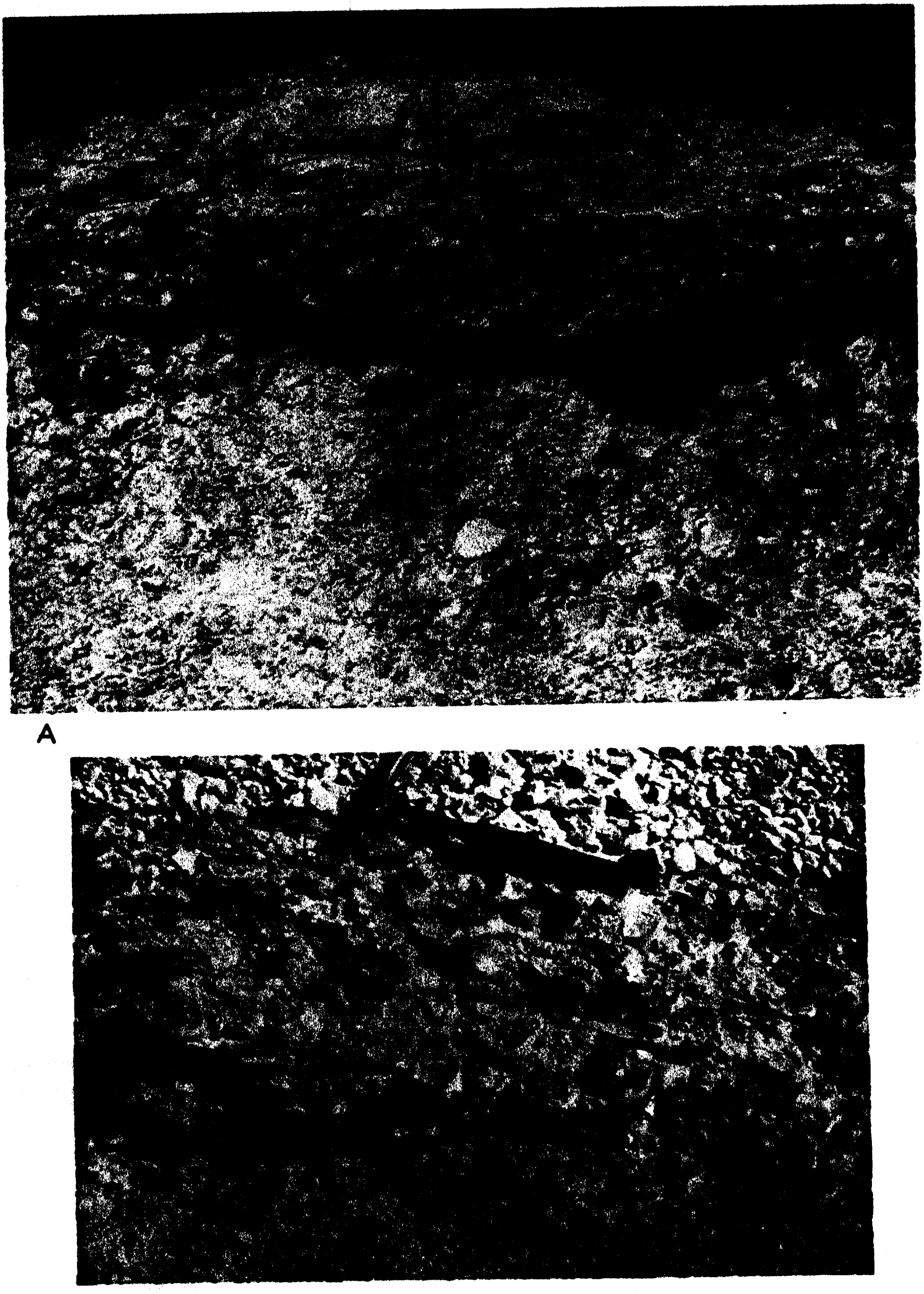
PLATE 4

A, Caliche fragment from surface rubble. A core of sandy caprock is surrounded by travertine bands. The present upper surface is outlined by solution, and undersurface has a recent crust of pustulose, spongy travertine. Hagerman Heights, ½ mile east of Carlsbad, N.M., along U.S. Highway 62. Specimen is 6 inches long.

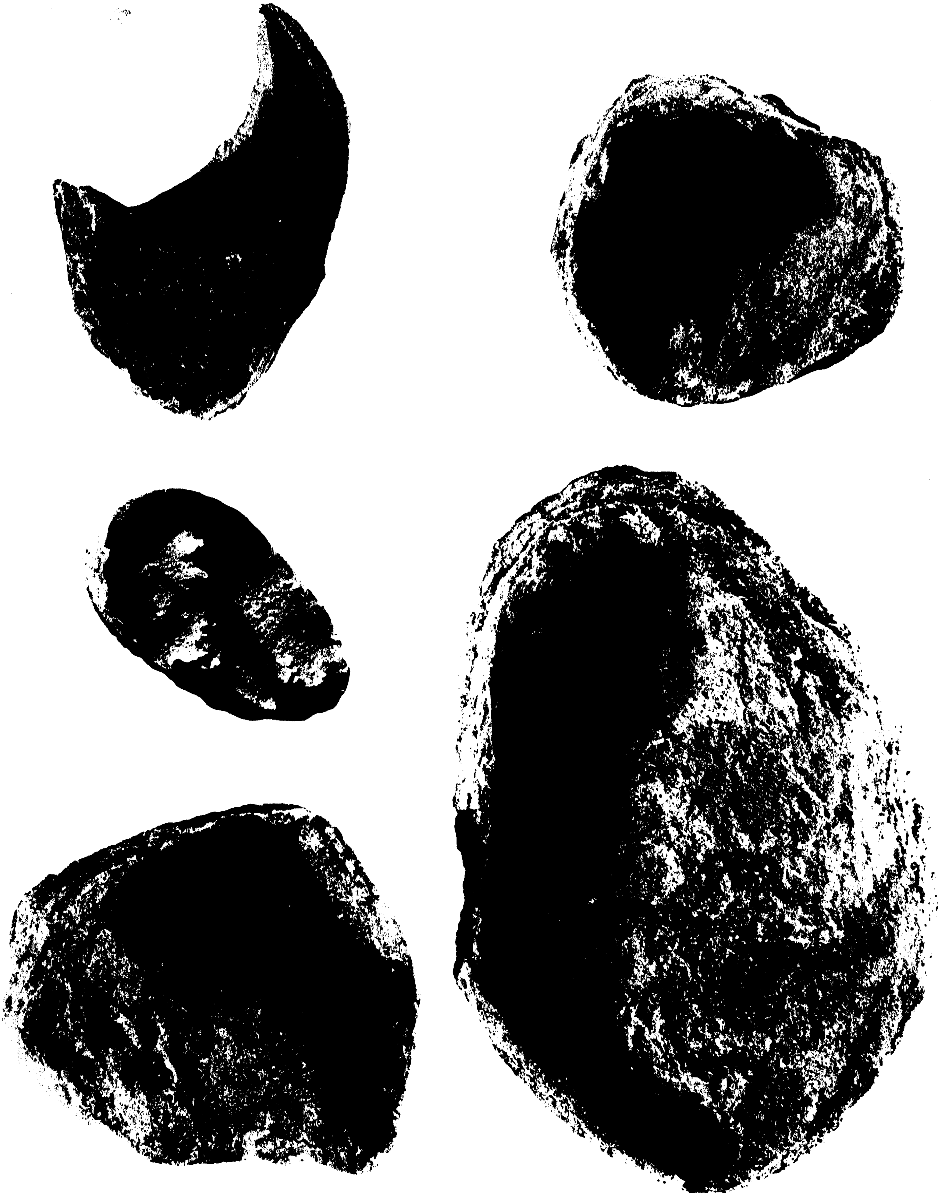
B, Degraded, spongy caliche below sandy soil. Solution cavities are filled with soil and caliche fragments. Junction U.S. Highway 62 and N.M. Highway 31, 2 miles northwest of Tower Hill (fig. 1).



Brecciated caprock and concretionary structure



Caliche profile



Cupped caliche pebbles

A

B

Caprock with recent travertine crust; degraded caliche and solution features.

caliche lies on all these surfaces except the Lakewood terrace, and in places it is found on transition slopes as well.

Limestone gravel constitutes the parent-material of the caliche at essentially all localities, and either it is present in the section, or its former presence is recorded by siliceous residuals. The gravels on the Llano and the Sacramento plain are of Ogallala age, whereas those on younger surfaces were derived largely from the Ogallala and from limestones in the Guadalupe and Sacramento ranges and are of Pleistocene age. Deposits occupying local basins on the Mescalero plain belong to the Pleistocene Gatuna formation (Robinson and Lang, 1939, pp. 84–85), and extensive dune sands on this surface are termed "Mescalero sands" (Darton, 1928*b*, p. 59).

The physiographic and geologic relations clearly indicate that the caliche varies in age from place to place and cannot be used as a stratigraphic horizon marker.

THE CALICHE PROFILE

DESCRIPTION

The two localities found in 1946 which best show the weathering profile of caliche overlying limestone gravel are near Hope, in northwestern Eddy County, and near Mesa, in southwestern De Baca County. Because both are in gravel-pit walls and will become obscured after the pits are abandoned, several other localities with less complete sections will also be noted.

The Hope Sections.—Two pits lie near N.M. Highway 83 east of Hope, in sec. 20, T. 17 N., R. 23 E., and sec. 17, T. 17 N., R. 24 E., respectively (figs. 1 and 3). Both are excavated in a partially cemented gravel of the Blackdom terrace along the Rio Penasco. Both possess a laminated and brecciated caprock a foot or so thick, the brecciation and rehealing indicated as having occurred two or three times (pl. 1, *A*). Below the caprock is a weak, chalky caliche grading down in 3 or 4 feet into limestone gravel (pl. 2). Few entire pebbles are recognizable in this zone. Those that are identifiable as originally pebbles possess an extraordinary amount of solutional reduction of the upper part. Original outlines are retained only on the lower half to three-fourths of the pebble surfaces. Tops are truncated by roughened, flattish new surfaces, clearly the result of solutional attack by down-moving soil water (pl. 3). In a considerable percentage of them the solutionally made surface is concave upward, with a sharp little rim bounding the concavity. In a few such pebbles the concavity has a surviving central nubbin, surrounded by a moat. In places the ensemble looks remarkably like a mass of clam shells, with all concavities upward.

The matrix of this zone is chalky and friable, and many pebbles may be picked loose with the fingers. Weak cementation also characterizes the gravel below the zone of this solutional modification.

Nye (Fiedler and Nye, 1932, p. 34) reported the pebble-faceting in one of these pits as due to rain-water solution during aggradation while each layer of pebbles successively lay on the surface. Such rain-water modification of limestone pebbles lying on the surface is a widespread phenomenon in the Pecos Valley, but there are serious objections to this interpretation for the solutionally truncated pebbles of the Hope sections.

Nye's hypothesis requires an exceedingly slow accumulation. This he recognizes. But it also requires the following improbabilities: (1) The ancient stream deposit must have lain dry for intervals of much more than a thousand years between the floodings that brought additional gravel; (2) each accretion of gravel was a layer only one pebble deep; (3) no overturning of already truncated and cupped pebbles ever occurred when a new layer was added.

Careful but fruitless search was made for an expectable relation, by Nye's hypothesis, of upper pebbles resting in the cups of immediately subjacent ones. Instead, several cases were found where a cavity lay immediately above a truncated pebble, the cavity arched over by laminated caliche matrix. In places where interstitial caliche rested directly on a pebble's truncated surface, it had a structure of short, close-set, vertical columns suggesting tiny stalactites and obviously was a deposit made in the cavity left by removal of the upper part of the pebble.

Although lamination is a common structure in the interstitial lime of this zone, it is limited largely to the lower sides of pebble remnants, and no case was found where laminated caliche was wrapped across the truncated upper surfaces.

It is concluded, therefore, that the 3-foot zone of altered pebbles is a weathering profile. Descending soil water has dissolved lime from the tops of pebbles during wet seasons or cycles. During succeeding dry spells, lingering films of this water, clinging to the bottoms of pebbles, deposited lime to make the laminated additions. The unaltered gravel beneath this zone may have lain below the water table while most of these changes were going on. Scattered entire pebbles of limestone found in the caprock must, by the writers' hypothesis, be later introductions.

A section in a large gravel pit along N.M. Highway 83 on the western outskirts of Artesia, 20 miles east of Hope, shows two different gravel deposits, both presumably post-Blackdom, separated by discontinuous red, laminated silt and clay. The top of the lower gravel has a thin layer of solution-truncated pebbles, both where covered by the silt and where the upper gravel rests directly on the lower. There is no accompanying calichification. Nye's explanation is perfectly appropriate here. A horizon of cupped pebbles in a gravel accumulation is a record of disconformity.

The Mesa section.—From Mesa to Vaughn, along N.M. Highway 285, the country rises from 4,500 to 6,000 feet A.T. through a series of erosional and sink-produced slopes which are not, like the Llano, remnants and records of a flat-topped depositional plain. Because, however, the slopes lie above the Blackdom level and carry many gravel residuals with siliceous pebbles, the fluvial materials are taken to be remnants of the Ogallala formation (Bretz and Horberg, 1949), and the surface is considered part of the Sacramento plain.

A large borrow pit for road material lies a little off the southwestern side of the highway, about 12 miles northwest of Mesa, at an altitude of about 4,700 feet. A.T. (figs. 1 and 2).[3] Being older than all terrace stages of Pecos Valley, it has expectably a more mature weathering profile (fig. 3). It does contain almost all the evidence necessary to prove that the caliche of southeastern New Mexico is the weathered product of limestone gravel.

The upper 1 foot of the section afforded by this pit consists of dense, slabby, banded, and brecciated caprock, with strong salmon-pink color and beautiful concretionary bandings to the center of

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[3] No more precise location has been obtainable.

the primary units of the breccia structure (zone *1*, fig. 2). Below this are 2 feet of less indurated caliche with less well-shown banding and more prominent breccia structure (zone 2). Texturally, it is still "Rock House" caprock.

A gradation of about 6 inches downward reaches a zone of nodular, poorly indurated caliche, 4 feet thick (zone 3). The nodules can be picked out of the face of the cut with the fingers. Cracked open, they reveal very good concentric structures.

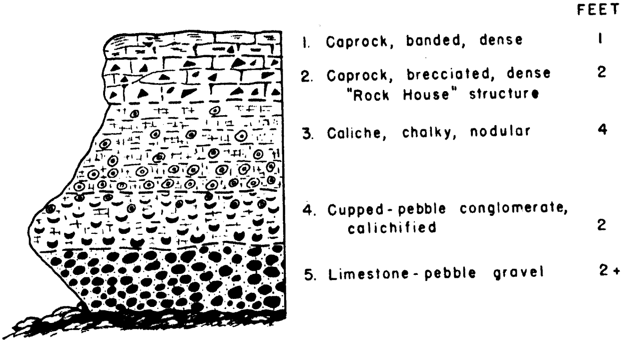
FEET
1. Caprock, banded, dense 1
2. Caprock, brecciated, dense "Rock House" structure 2
3. Caliche, chalky, nodular 4
4. Cupped-pebble conglomerate, calichified 2
5. Limestone-pebble gravel 2+

FIG. 2.—Zones in the caliche profile. Caliche pit on Sacramento plain, 12 miles northwest of Mesa, N.M. "Rock House" structure is a field term for the breccia, being named from caprock specimens from Rock House, N.M.

From 6 inches to 1 foot deeper, many of the nodules have a cup on the upper side.

At a total depth of 6–8 feet these cupped nodules consist of caliche layers only for their lower part. The smaller, upper part is a remnant of a limestone pebble!

A foot or so lower in the section, the remains of the limestone pebbles exceed in bulk the caliche part of the nodules (zone *4*). The limestone is soft and weathered.

A little lower, pebbles take the place of nodules, all cupped on the upper side. The limestone is bright and unweathered in the center, and there is only an incipient caliche coating on the bottom. The lower 2 feet in the section consist of unweathered limestone-pebble gravel (zone *5*).

Quartzose pebbles occur throughout the entire section.

The conclusion drawn from this section is that the entire sequence was once a limestone gravel with scattered siliceous constituents and is now a weathering profile showing greatest alteration at the top, where it has developed the typical caprock structure, a surface-weathering product of this region's climate on parent-material of limestone gravel. The gradation downward into nodular caliche suggests that primary units of the breccia caprock may owe their origin to such nodules. The gradation of nodules into pebbles indicates nodule origin from wastage of upper surfaces and accretionary growth on lower surfaces of original pebbles. Thus each nodule is a record of a former pebble. So complete is this transformation in the upper part of the nodule zone that many are concentric to the core. Others still enclose a remnant of the original limestone pebble or may hold a quartzose pebble. As nodule growth beneath a wasting pebble has progressed, it

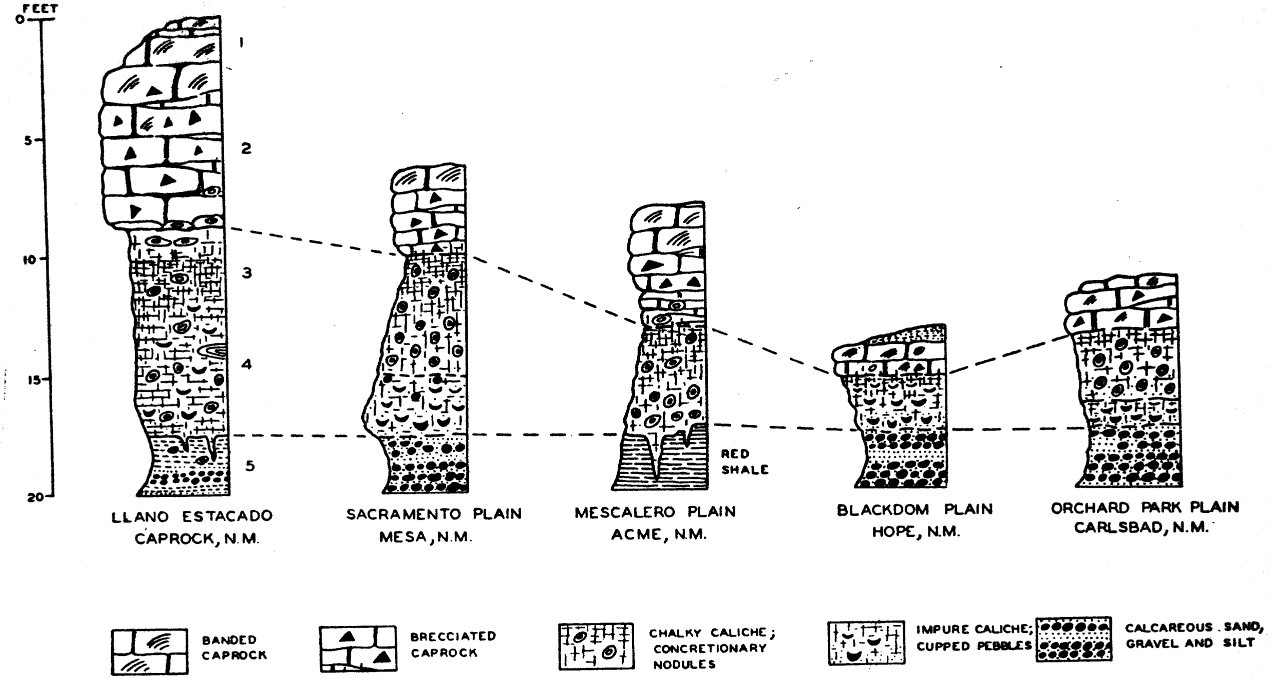

FIG. 3.—Variation in caliche profiles on surfaces of different ages. Llano scarp near Caprock, N.M., 1½ miles south of U.S. Highway 380; caliche pit about 12 miles northwest of Mesa along U.S. Highway 285; caliche pit, 2 miles southwest of Acme, N.M., along U.S. Highway 70 (23 miles northeast of Roswell, N.M.); caliche pit, 1 mile northeast of Hope, N.M.; cut along U.S. Highway 62 at Hagerman Heights Carlsbad, N.M.

has come to include small bits of other material, perhaps also concretionary in structure, perhaps quartzose.

The one difficulty with the writers' interpretation is that, where most of the nodules should show, in their cross sections, a series of diminishing, concave-upward saucer-forms toward the top, most of them show an equidimensional concentricity. Some, however, show the structure expectable by the genesis proposed.

A similar section, but with the cupped-pebble zone less well shown, is exposed in a caliche pit along the same highway, about 5 miles farther northwest.

Hagerman Heights section.—Just east of Pecos River at Carlsbad is a definite terrace known as "Hagerman Heights." It is considered a part of the Orchard Park surface (fig. 1). Grading for U.S. Highway 62 has made a long cut through the scarp of this terrace, and in the sides of this cut is an instructive section (fig. 3).

Caprock is wholly limited to the flat surface of the terrace. It averages 18 inches in thickness but locally is 4 feet thick. It possesses concentric banding of laminae around older caliche fragments and embedded quartzose pebbles. Manganese oxide "ink lines" are involved in this banding.

Below the caprock are 3–4 feet of rather loose, nodular, chalky, poorly indurated, and poorly layered caliche, with but few quartzose pebbles as compared with the caprock. Their relative abundance in the caprock can well be the result of concentration through solutional removal of associated calcareous pebbles.

This nodular caliche grades down into about a foot of gravel, in which many limestone pebbles have had their upper portion eaten away by solution *in situ*, leaving "half-shell" or cupped forms, comparable to those in the Hope and Mesa sections.

Below this zone is a 5-foot bed of cemented limestone gravel with pebbles entire and, below that, a brown stratified silt with incipient calcareous cementation.

Telltale Bluffs section—A cliff, continuous for a mile along the east side of Pecos River, about 2 miles southeast of Carlsbad, is called "Telltale Bluffs." The basal part of the cliff is of strongly tilted Permian limestone beds, above which are 50 feet or so of the "quartzose conglomerate" formation (Bretz and Horberg, 1949), composed chiefly of Permian and Triassic red-bed debris but with siliceous pebbles in every stratum. The amount of red-bed contribution decreases toward the top, giving way to a dominance of grayish, largely limestone conglomerate and lime sandstone. The upper 10 feet of this may be described as a nodular aggregate, and the top 3 feet or so are typical caprock, indurated enough to overhang the cliff brink for the entire mile of length. The summit of Telltale Bluffs lies in the Orchard Park niveau.

The caprock breaks into slabs along nearly horizontal planes. Quartzose pebbles are scattered throughout. The texture is fragmental, older pieces of the caprock being surrounded by concentric laminae and those concretionary units being embedded, although not in contact, in a dense caliche matrix. Where the original fragments were small, the effect is that of a rude pisolitic texture.

In the northern part of the bluff the nearly horizontally layered caprock overlies and truncates a 15° dip in the underlying conglomerate and sandstone. This disconformity is interpreted as a weathering profile superposed on the underlying highly calcareous conglomerate, and the stratiform structure of the caprock is interpreted as developed *in situ* by this process rather than being true strata of successive deposition on the underlying tilted rocks.

Inconspicuous but consistently present, at depths of 8–10 feet, are cupped pebbles. Above their level is nodular caliche, below it the limestone pebbles are entire.

Another river-bluff section, about 1 mile south of Telltale, shows a feature seen in several places in caliche immediately subjacent to caprock. It is that of "eviscerated" pebbles, the solutional attack having first breached the upper surface and then having dissolved away much of the interior, leaving a hollow partial shell, which later has become filled with sandy material. This, however, may have occurred during aggradation of the deposit rather than during the later calichification.

Carlsbad airfield section.—A large roadside pit for highway material, excavated in gravel of the Orchard Park plain, lies across U.S. Highway 62 from the airfield, 2 miles south of Carlsbad. The caprock is the familiar caliche breccia, cemented by regrowth to a dense, slabby material. Below this are about 6 feet of nodular, chalky, punky caliche containing a few siliceous pebbles and some still recognizable limestone-pebble outlines.

Gradation from this nodular material down into firm limestone conglomerate with its pebbles entire is by an irregular zone of truncated pebbles, with numerous cupped forms.

Exposures between Lovington and Maljamar, Lea County.—West from Lovington, along N.M. Highway 83 to the Llano scarp overlooking the Mescalero plain, there are numerous shallow cuts and scrapings in the surface of the caprock, which show what is believed to be genetically a significant feature of the rock.

The Llano plain along this route is an almost flat surface, diversified only by a few broad, shallow sinks. Caliche caprock lies beneath a thin soil composed in part of wind-deposited material. Where exposed, the caprock is a dense limestone but is readily cracked and broken into slabs by road-making machinery. It parts along "bedding" planes and already existing vertical cracks. Rarely is any movement of slabs from position shown, except where disturbed by the grader.

Texturally, each of these slabs is a breccia aggregate. The internal arrangement of component parts shows no tendency whatever toward a stratification paralleling top and bottom of the slab. The structure of the fragments constituting the breccia shows that they have been derived from an older laminated porcellanous caliche.

The fragments are rarely in contact. They are characteristically surrounded by layers of later-deposited calcium carbonate. Some also possess thin rims of manganese dioxide. The matrix embedding the fragments is a sandy, fine-textured lime. The aggregate is so similar to that shown in plate 1, *A*, and figure 5, from Rock House, New Mexico, that, without proper labeling, even the collector would be at a loss to differentiate.

Before being disturbed by the grader, these slabs were already discrete units. They can be pried loose with a hammer. The upper surfaces of slabs are covered by porcellanous caliche, finely laminated parallel to that surface but with the laminae turning over the edges to descend vertically into bounding cracks. The bottoms of the slabs are characterized by a pustulose surface of added lime carbonate, commonly pinker in color

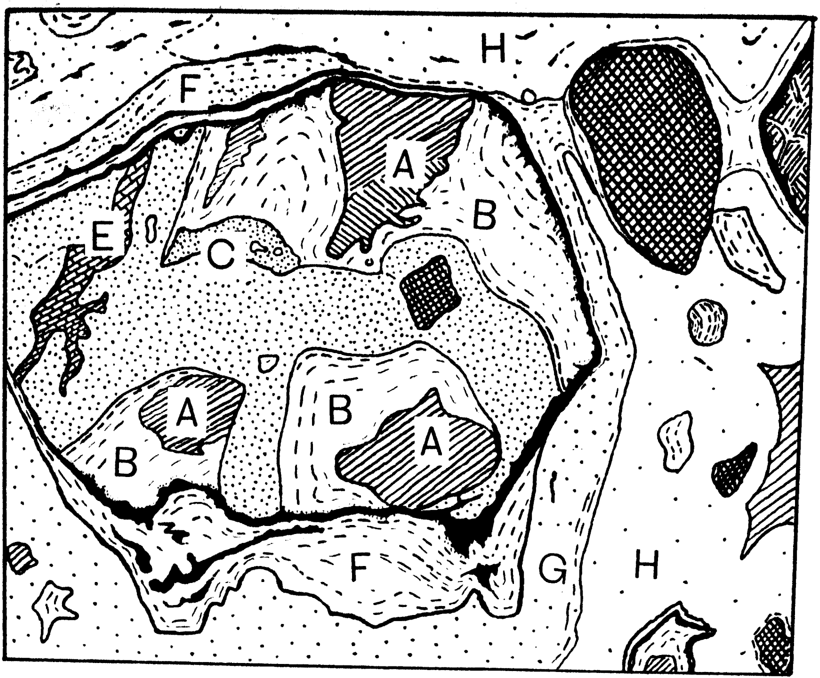

Fig. 4.—Polished surface of brecciated caprock from Rock House, N.M., showing several generations of fragments and matrix. *A*, crystalline travertine; *B*, sandy, banded travertine; *C*, sandy caliche; *D*, sandy matrix; *E*, travertine veinlet; F, banded travertine; *G* and *H*, sandy matrix. Siliceous pebbles are cross-hatched; manganese oxide rims are in solid black. Width of surface is 45 mm.

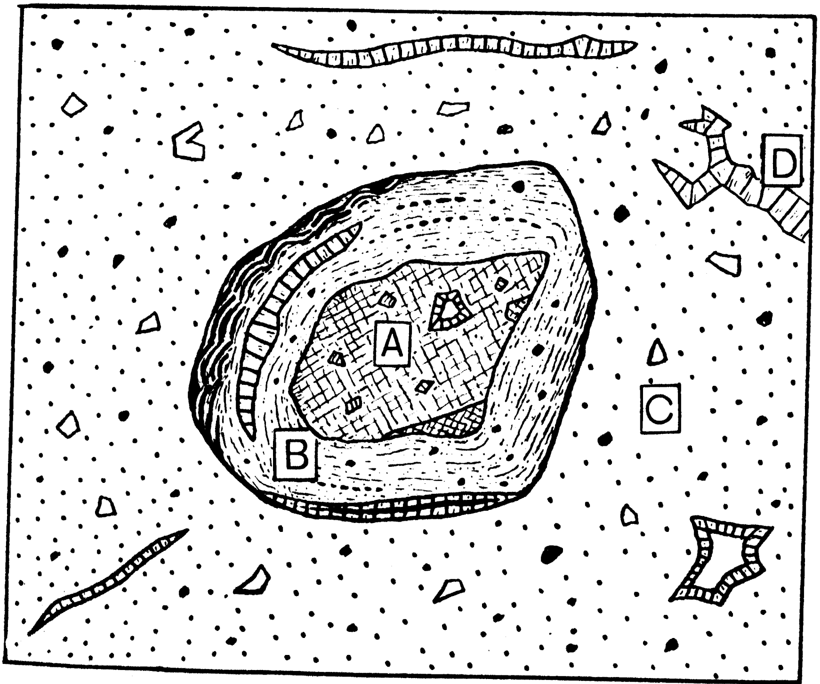

Fig. 5.—Thin section of brecciated caprock from Rock House, N.M. *A*, fine crystalline travertine; *B*, banded travertine; *C*, sandy matrix; *D*, coarse crystalline travertine veinlet. Manganese oxide rim in solid black. Width of section is 6 mm.

than the surface laminae but of just as dense and porcellanous material. This bottom addition does not share in the concentric wrapping of the slab. It truncates the lower edges of the laminae on the slab margins and appears to have no laminae in itself. Truncation appears to be due to development of solutional surfaces before deposition of the bottom crust. Solution markings are common on upper surfaces of the slabs.

Clearly, these slabs have been growing *in situ;* the pustulose underpart may still be growing. The writers believe that these additions to the slabby fragments of the brecciated caliche have taken place on the surface of the Llano, just under a thin, noncalcareous soil, through the agency of rain and soil water. The added material is believed to have come from wastage of once overlying portions of the original caliche, whose basal part now survives in the slab interiors.

Were these slabs to become more broken by weathering or by deposition in the cracks and the more nearly equidimensional pieces rotated and then recemented to become a continuous rock body again, another episode of brecciation would be added to the caprock's history. Such repeated cycles of brecciation, solutional wastage, and recementation by growth of laminar wrappings is strikingly recorded in the specimens from Rock House (pl. 1, *A;* and fig. 4).

Although showing only the very top of the caliche profile, these exposures west of Lovington are miles from any river bluff or terrace scarp, and they demonstrate that caprock antedates the declivities at whose summits it may crop out and therefore antedates the groundwater levels influenced by those declivities.

The profile of weathering below the caprock in this general area is revealed by sections at the Llano scarp near Maljamar. In a cut along N.M. Highway 83, 1 mile east of Maljamar, a thin remnant of the caprock is underlain by about 15 feet of chalky, porous caliche in rude layers, 1–3 inches thick, and contains lenticular masses, 2–6 feet in length, of dense brecciated caliche similar to the caprock (pl. 1, *B* and *C*). This grades downward into the parent-material—a brown, silty, calcareous sand of the Ogallala formation. A similar section is exposed along N.M. Highway 31, about 12 miles to the northwest, where 10 feet of dense, brecciated caprock are underlain by an equal thickness of chalky, concretionary caliche, which contains lenticular masses of dense caliche up to 6 feet in length. The underlying Ogallala formation is a fine, silty sand. A partial section of the Llano's weathering profile may be seen in a caliche pit along N.M. Highway 83, about 11 miles east of Maljamar, where the layered structure and lenticular masses in the chalky caliche zone are particularly well shown. Siliceous pebbles appear to be absent at all these localities. The sections differ from those previously described by the absence of a cupped-pebble zone and a well-defined nodular zone. This appears to be due to the fine-grained texture of the parent-material. The thicknesses of the caprock and chalky zone is greater than that on younger surfaces and typifies a mature stage of development.

Section near Caprock, Lea County.—Another mature caliche profile on fine-grained Ogallala sediments is exposed at the Llano scarp, about 5 miles southwest of Caprock (fig. 3). In this section the caprock reaches a thickness of about 30 feet and is composed of dense, massive, laminated, and brecciated caliche which contains relicts of nodular structures. It grades downward into chalky, nodular caliche, which is about 10 feet thick and, like that near Maljamar, is rudely layered and contains dense, lenticular masses, 1–2 feet in length. Inconspicuous small chert nodules and stringers occur in the chalky caliche. The unaltered Ogallala deposits at the base of the section consist of sandy silts with pebble bands and calcareous nodules and veins. Siliceous pebbles occur throughout the section. Some of the nodules in the chalky zone resemble altered limestone pebbles, but definite identification was not possible. The abundance of nodules, however, suggests that limestone gravels were once present and were destroyed during calichification.

Section near Taiban.—A thin caliche profile developed on Ogallala sand is exposed at the Llano scarp, 3 miles north of Taiban, De Baca County, and is repeated along the scarp at localities to the east and southwest (fig. 1). The caprock is only 2–3 feet thick and has a chalky appearance. The underlying porous caliche, which is about 3 feet thick, is rudely layered and contains scattered nodules of dense brown travertine. This grades downward into a silty, calcareous sand, about 15 feet thick. There are calcareous veins and nodules in the upper part, and the lower 3–4 feet are cemented into a mass of nodular, concretionary vertical pipes. Underlying the sand is a coarse, poorly indurated conglomerate with boulders up to 14 inches in diameter. Although the thinness of the profile, as compared with other Llano sections, may be due in large part to removal of the caprock by erosion, it is also possible that it reflects a lower calcium carbonate content in the parent-material.

Yeso section.—A caliche section on an outlier of the Llano plain west of the Pecos is exposed in a pit along the north side of U.S. Highway 60, 7 miles northwest of Yeso, De Baca County (fig. 1). At this locality a dense, brecciated, and laminated caprock, 3 feet thick, overlies about 5 feet of soft, nodular, rudely layered caliche. This, in turn, rests on and penetrates downward into brecciated Triassic red beds without intervening Ogallala sediments. Siliceous pebbles in the caliche, however, indicate that Ogallala gravels were once present. Nodules and fragments in the soft caliche zone are composed of dense salmon-colored travertine similar to the caprock, and many are solution-faceted and encrusted with secondary caliche. These relations suggest a second-cycle profile in which earlier caprock formed the parent-material. It is probable that this process was continuous during lowering of the surface and that the original parent calcareous sand and gravel may have lain well above the bedrock now being intruded by caliche.

Acme section.—Another second-cycle profile is shown in a caliche pit along U.S. Highway 70, about 22 miles northeast of Roswell (2 miles southwest of Acme, Chaves County) (fig. 3). The caliche in this area forms the surface of the Mescalero plain, and east of Acme it is covered by Mescalero dune sands, which are probably of Pleistocene age. The section consists of dense, laminated, and brecciated caprock, 3 feet thick, underlain by 6 feet of soft nodular caliche, which overlies and penetrates downward as caliche roots into Triassic red beds. Siliceous pebbles are present in the caliche. The nodules are composed of dense travertine and are encrusted with layers of softer caliche.

CONCLUSIONS REGARDING THE CALICHE PROFILE

The weathering profile in the calcareous gravels of the Ogallala formation of the Llano and three successively younger and lower surfaces between the Llano and the mountains to the west consistently possesses a caprock of (1) laminated caliche and (2) firmly cemented caliche breccia; (3) a subjacent zone of crumbly, nodular, chalky caliche with no pebble outlines, below which there is gradation by (4) a zone of cupped pebbles into (5) limestone gravel unaltered except by cementation (fig. 2). Exposures vary; but, if deep and clean enough, one may expect to find evidence for these alteration units wherever calichification has occurred on a limestone gravel in the region.

Except close to the Guadalupe and Sacramento limestone ranges, whence came the abundant calcareous gravel of the Ogallala formation, one may anticipate finding scattered siliceous pebbles in all four altered units, as also in the deeper unaltered gravel. Their occurrence in the brecciated caprock was as much a puzzle during the early weeks of the field study as was the compounded breccia itself. It is confidently asserted that they will remain an unsolved puzzle for anyone who refuses the explanation here advanced.

A comparison of caliche profiles on surfaces of different ages (fig. 3) shows that there is a general increase in the depth of the profile with age. There appear to be many exceptions, however, as indicated by the Taiban section, which is abnormally shallow, and the section at Hagerman Heights near Carlsbad, which is abnormally deep. Many additional measurements would be needed to establish this general relation on a quantitative basis.

Age is but one factor in determining the depth of the weathering profile. Topographic setting, rainfall, erosional attack by wind and water, temporary burial by eolian sand, and the composition and texture of the parent-material—all these and perhaps other factors have functioned. But downward migration of the bottom of the nodular zone as pebbles have been consumed has been continuous, and the front of this advance has been marked by the zone of cupped pebbles. The Hope and Mesa sections provide a detailed record of the transitional stages between limestone gravel and nodular caliche.

In the compound breccias of the caprock is the record of climatic oscillations, alternating between pluvial and semiarid during the history of the calichification.

In the laminae locking the breccia fragments together is the record of solutional wastage and disappearance of somewhat higher levels of the original calcareous gravel or sand deposit.

Where caliche overlies noncalcareous red beds, the writers are persuaded that the original characters of an original limestone gravel or sand have all been destroyed in calichification.

Caliche on marine limestone bedrock has been found only in proximity to a calichified limestone gravel on this bedrock and appears to record the thin edge of that gravel cover when the alteration took place. A proper climatic environment and a calcareous substratum do not seem to encompass all the requirements for caliche-making. The abundant interstices of sand and gravel appear to be necessary, also.

LITHOLOGIC DESCRIPTION

POLISHED SURFACES AND THIN SECTIONS

Caprock.—Microscope studies show that the varying textures in both banded and brecciated caprock are due to differences in (1) amounts of detrital impurities, (2) crystal grain size of the caliche, and (3) manganese and iron oxide colorations. Although a sharp boundary cannot be drawn between the two caprock zones, the upper banded caprock is more massive and crystalline and has fewer impurities.

The characteristic texture of the caprock is best displayed in the brecciated zone and consists of unoriented fragments of caliche in a sandy limestone matrix (pl. 1, *A;* fig. 4). Individual fragments are composed of banded crystalline travertine, older sandy matrix, concretionary nodules, siliceous residuals, or masses of older breccia. In places there is evidence of partial replacement of fragments by the matrix. Several generations of breccia are shown. Many of the fragments are surrounded by layers of concentrically banded travertine and are rimmed by an impure black manganese oxide glaze. The glaze defines the fragments as sharply as if they were outlined in ink and is one of the most striking features of the rock. It margins solutional surfaces and penetrates irregularly into the fragments. Chemical analyses of the glaze show a concentration of manganese oxide, but no higher percentages of iron oxide than are present in the associated caliche.

As far as determinable, the carbonate in the caprock is calcite. No aragonite was noted, and chemical analyses show only a small percentage of magnesium oxide. Siliceous fragments and pebbles in the caprock are residual from parent-limestone gravels and consist of quartz, quartzite, chert, and jasper in about that order of abundance.

A complex history of brecciation, solution, precipitation, and cementation is recorded by almost all caprock sections. The polished surface shown in figure 4 is typical and indicates the following sequence of events: (1) brecciation of crystalline travertine *A* and solution of the margins of fragments; (2) deposition of sandy banded travertine *B;* (3) brecciation (?) and solution; (4) deposition of sandy caliche *C;* (5) brecciation and solution; (6) cementation by sandy matrix *D;* (7) formation of travertine veinlet *E;* (8) brecciation and solution; (9) deposition of manganese oxide glaze; (10) deposition of brown banded travertine *F* and associated manganese oxide streaks; (11) solution; (12) deposition of sandy matrix *G;* (13) brecciation and solution; and (14) cementation by sandy matrix *H*. A similar sequence, though less complicated, is shown by the thin section in figure 5. The thin section also emphasizes textural contrasts primarily due to differences in impurities and crystallinity and the fracture outlines of calcite veinlets and vugs.

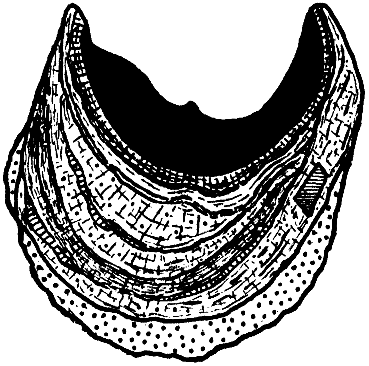

FIG. 6.—Nodule from chalky caliche zone with cupped limestone pebble (*solid*) enclosed in travertine crust. The travertine layers represent differences in purity and crystallinity. The inner band is darker reddish-brown, dense travertine; the outer crust is a porous, chalky caliche. Width of specimen is 1.5 inches. Caliche pit, 2 miles southwest of Acme, N.M., along U.S. Highway 70.

Chalky caliche.—Textures and structures in the chalky caliche below the caprock are not so well defined as are those above. The nodules in the upper part of the zone have cores of banded or dense sandy caliche similar to that of the caprock and are enclosed by concentric laminae of travertine. Commonly the outer crust is tuffaceous and chalky, and the innermost band is a dense brown travertine. In most specimens there are inclusions of older caliche and solutional surfaces between travertine crusts. In the lower part of the zone the nodules give way to solution-cupped limestone pebbles and nodules encased in travertine crusts which are thickened at the bottom (fig. 6). The cupped nodules, which lie near the top of the cupped zone, indicate an overlap in processes of cupping and nodule development. A porous, punky, impure caliche forms the matrix and extends downward into the parent-limestone gravel.

TABLE 1

WEIGHT PERCENTAGES OF ACID-INSOLUBLE RESIDUE IN CALICHE OF NEW MEXICO

Caliche caprock:	
Llano Estacado	
1. Lovington	Trace
2. Maljamar	1.8
3. Kermit	2.9
4. Santa Rosa	3.1
5. Yeso	2.1
Sacramento plain	
6. Vaughn	2.1
7. Mesa	2.4
Diamond A plain (?)	
8. Chaves	Trace
Mescalero plain	
9. East of Roswell	3.0
10. Rock House	3.8
Blackdom plain	
11. Hope	Trace
Younger surface	
12. Carlsbad	8.7
Average	3.33
Nodular caliche:	
13. Llano at Caprock	39.8

INSOLUBLE RESIDUES

The weight percentages of residues insoluble in dilute acid in the caprock range from a trace to 8.7, and in nine measurable samples the average was 3.33 (table 1). In one sample from the nodular zone the percentage was 39.8. Most of the residue is fine sand and silt, with an average estimated percentage of 85.4 quartz particles. Polished and pitted grains were noted in four samples, and rounded grains were prominent in eight samples out of the thirteen. Minor percentages of the residue consist largely of chert, chalcedony, dark minerals, iron oxides, and clay-size particles. There is no indication of volcanic ash, such as was reported from residues of caliche near Lubbock, Texas (Sidwell, 1943). More than half the quartz grains are angular, and many are subhedral, indicating a derivation from weathering of crystalline rocks. The possible importance of mechanical weathering during calichification is suggested by the dominance of silt and fine sand-size particles.

ALTERATION OF CALICHE

At most localities in southeastern New Mexico there is evidence that the caliche formed under earlier climatic regimes and has undergone extensive alterations in Recent time. This evidence consists of solution features, degraded caliche profiles, relations to present soils, and surface alterations of the caprock. Other alterations include structures due to expansion and development of concretionary masses.

SOLUTION FEATURES

Irregular solution cavities penetrating downward into the caliche are common features where there are overlying noncalcareous soils or unconsolidated deposits (pl. 4, *B;* figs. 7 and 8). The cavities vary in length from a few inches to several feet and are filled with the overlying deposits mixed with caliche fragments. In many places the cavities are lined with a thin crust of crystalline travertine. The shapes of some of the cavities (pl. 4, *B*) suggest that they may have been determined originally by roots and in some cases by animal burrows. Relict solution cavities and fills were noted at several localities and indicate the cavities and of zones of caliche fragments within the fill. Unaltered, dense, brecciated caprock forms bosses on the floor of the pit. The shapes and dense composition of nodules and fragments in the fill and in the degraded caliche indicate differential solution of original caprock structures.

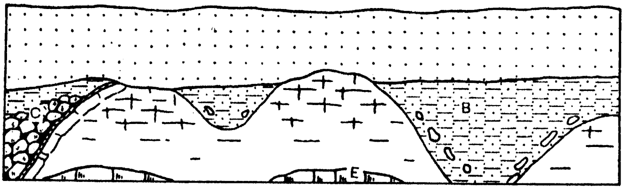

FIG. 7.—Solution cavities in caliche pit, ½ mile south of Kermit, N.M. *A*, reddish-brown, sandy and silty alluvium; *B*, red silt with caliche fragments; *C*, caliche fragments and nodules in travertine-lined cavity; *D*, chalky caliche with rude layers; *E*, dense travertine caprock. The exposure is about 10 feet high.

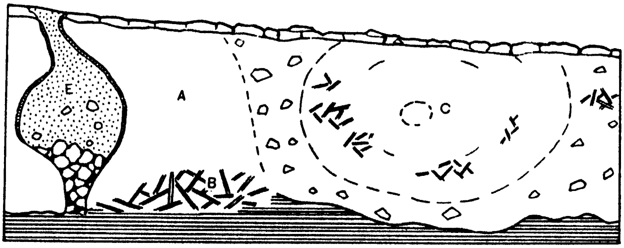

FIG. 8.—Solution cavity and concretionary structure in road cut, 1 mile south of Tower Hill, N.M. *A*, buff, chalky caliche; *B*, breccia of fragments of underlying Permian red beds cemented with caliche; *C*, dense, concretionary caliche with relict breccia of red bed fragments; *D*, brecciated caliche; *E*, solution cavity with travertine lining, filled with soil and caliche fragments. The exposure is 8 feet high.

that at least some solution preceded the latest stages of calichification.

Features due to solution, slumping, and filling are well shown in a caliche pit on the Llano near Kermit, where decomposed caliche is overlain by dune sand and slope-wash deposits (fig. 7). Broad, funnel-shaped cavities lined with travertine descend below the level of the pit floor and are filled with residual red silt and caliche fragments and nodules. Slumping due to solution is evidenced by steep dips of the caliche at the margin of

Solution features on a larger scale, represented by basins and broad shallow valleys, are well known on the Llano caprock (Price, 1940*b*) but are less conspicuous on younger surfaces.

The caliche associated with the solution features is commonly soft and porous, as compared with sections in which overlying soils and surficial deposits are absent. Its derivation from normal caliche is indicated by its gradation downward into unaltered caliche and by relict structures of the caprock and nodular zones. In some caliche sections, where solution is not clearly evidenced, it is possible that the porous caliche represents a normal transition into the *B* zone of the caliche profile.

RELATION TO PRESENT SOILS

Numerous exposures in the area, especially on the Llano and the Mescalero plains, show gradation of present soils downward into caliche and indicate that the caliche is the parent-material from which they were derived. A typical section on the Mescalero plain shows a downward gradation of (1) a thin organic zone, (2) silty sand with calcareous veins and nodules, (3) chalky nodular caliche, and (4) dense caprock. This condition appears to exist over wide areas of the southern High Plains, and pedologists regard caliche as parent-material for many soils in the region (Marbut, 1923, p. 66; Hawker, 1927; Harper and Smith, 1932, pp. 18–19). Many other soils have been developed from younger sediments deposited on the caprock (Harper and Smith, 1932, p. 16).

SURFACE ALTERATIONS

Where the caprock is exposed at the surface, as is the case over wide areas, it is broken into angular blocks and constitutes a surface rubble. The upper surfaces of these blocks are cupped and faceted by solution, and their sides and bottoms are encrusted with later additions of a pustulose, spongy caliche (pl. 4, *A*). These features are clearly the results of well-known processes of solution-faceting (Bryan, 1929) under present climatic conditions. Continued for an adequate time under sufficiently vigorous climatic conditions, such fragmentation and solutional wastage on the surface with accompanying deposition of subsurface crusts in cracks should produce the banded and brecciated texture exhibited by caprock caliche.

EXPANSION STRUCTURES AND CONCRETIONARY MASSES

Deformed layers and breccias due to expansion of the caliche or underlying shales are known at several localities. In places this is accompanied by development of dense caliche masses within the chalky caliche zone. The expansion appears to have resulted largely from weathering and impregnation of calcium carbonate during later stages of calichification. It is probable that a considerable proportion of the breccia of the caprock is due to earlier action of this process as well as to desiccation.

Lenticular masses of dense, banded, and brecciated caliche up to 10 feet in length occur in the Ogallala caliche and represent the "algae reefs" described by Price, Elias, and Frye (1946). Several masses well down in the chalky caliche zone are exposed along N.M. Highway 83 at the Llano scarp, 1 mile east of Maljamar (pl. 1, *B*, *C*). Rude layers in the enclosing caliche curve irregularly around the lenticular masses. But this enclosing laminated caliche may also pass through the enclosed central mass in a way that demonstrates consanguinity with the laminae wrapping around tops and sides of caliche slabs along N.M. Highway 83, east of Maljamar. Such crossings of the larger mass are crack-fillings, not organic precipitates; and unbroken continuity of their laminae with those enclosing the central mass is demonstrable. The structure of the central mass is that of a dense, laminated, travertine breccia similar to the caprock. Similar masses occur in a caliche pit along the same highway, about 11 miles to the east, and along N.M. Highway 31 at the Llano scarp, 12 miles northwest of Maljamar. Ovate masses of dense caliche containing large angular fragments were also noted farther north in a caliche pit, 1½ miles west of Kenna (fig. 1). At this locality deformation of the enclosing softer caliche by expansion is evidenced by low domes on the floor of the pit. The upper surfaces of the domes are marked by septarian-like fractures filled with banded travertine.

A spherical mass of less dense caliche with concentric fractures was noted in younger degraded chalky caliche on the Mescalero plain, 1 mile south of Tower Hill (figs. 1 and 8). This structure is clearly concretionary and is surrounded by a zone of breccia, evidencing expansion during its growth. It suggests an early stage of concretionary growth which might later give rise to the reef-like masses described above.

ORIGIN OF CALICHE

GENERAL SUMMARY

Calcareous caliches appear to be a world-wide feature of arid and semiarid regions, and a formidable literature by pedologists, geologists, and geographers could be assembled. Caliche or calcareous surface crusts and hardpans are reported from the Great Plains south of Nebraska and throughout the Southwest (Blake, 1902; Lee, 1905, p. 107–111; Marbut, 1923, p. 66; Udden, 1923; Gould and Lonsdale, 1926, pp. 29–33; Trowbridge, 1926; Hawker, 1927; Breazeale and Smith, 1930; Elias, pp. 136–141, Price, 1933; Sayre, 1937, pp. 65–72; Schoff, 1939, pp. 79–83; 1943, pp. 100–113; Smith, 1940, pp. 44–45; Frye, 1945, pp. 89–91; Price, Elias, and Frye, 1946); Mexico (Price, 1925); Argentina (Kashirsky, 1938); South Africa (Du Toit, 1939, pp. 407–408); Libya, Tunis, Morocco, Algeria, Syria, Egypt, Palestine (Glinka, 1927, pp. 157–159); Arabia (Miller, 1937); and Australia (Woolnough, 1928*a;* 1928*b;* 1930).

Numerous theories have been proposed for the origin of caliche-like materials, and it is unlikely that any one theory is of general application. Most of the theories proposed are included in one or more of the following groups:

1. Deposition in extensive shallow lakes by algae and inorganic processes, as proposed for the Ogallala caprock by Elias (1931, p. 141; 1948, p. 611)
2. Deposition in small disconnected lakes and ponds by physical and/or organic processes (Theis, 1936; Frye, 1945, pp. 91–95)
3. Deposition along streams, especially intermittent streams, by physical and/or organic processes (Trowbridge, 1926; Price, Elias, and Frye, 1946)
4. Deposition by rising artesian waters either at the water table or at the surface (Blake, 1902)
5. Deposition by capillary rise of water from the water table, especially under conditions of high water table (Lee, 1905, pp. 107–111) or a rising water table brought about by aggradation (Theis, 1945)
6. Deposition in the *B* zone of the soil profile by descending surface waters or by capillary rise of surface waters following saturation of the soil zone (R. H. Forbes, quoted in Lee, 1905, p. 110; Glinka, 1914, pp. 157–159; Marbut, 1923, p. 66; Udden, 1923; Gould and Lonsdale, 1926, pp. 29–33; Hawker, 1927; Price, 1933, 1940; Sayre, 1937, pp. 69–70)
7. Various combinations of the above (Lee, 1905, pp. 107–111; Trowbridge, 1926; Breazeale and Smith, 1930; Price, Elias, and Frye, 1946)

All these theories have been applied to caliche in the High Plains and the Southwest, and all except the artesian theory have had recent proponents.

ORIGIN OF CALICHE IN SOUTHEASTERN NEW MEXICO

General statement.—Any satisfactory explanation of the caliche in southeastern New Mexico must account for the following features: (1) the profile development recorded by caliche zones; (2) variations in the profile determined by parent-material and physiographic position; (3) presence of siliceous residual pebbles throughout the profile; (4) caliche of various ages on four different physiographic surfaces and on intervening slopes; and (5) fundamental similarities in the profiles and textures in caliches of various ages; (6) continuity and wide lateral extent, especially on the Llano surface; and (7) occurrence of concretionary and banded "algae-like" structures.

Lacustrine hypotheses.—The lacustrine hypothesis for the Ogallala caprock, as first presented by Elias (1931, pp. 136–141), proposed that the caprock differed from true caliche and was deposited largely by algae in extensive shallow lakes on the surface of the High Plains. Algae cells belonging to the genus *Chlorellopsis* were identified from banded structures in the caprock. Later the caprock was traced southward as an essentially continuous "key bed" into the Llano of southeastern New Mexico, and algal reefs as well as smaller algal structures were described (Price, Elias, and Frye, 1946). Modifications of the hypothesis, restricting the lakes to small, discontinuous bodies of water, were proposed by Theis (1936) and Frye (1945, pp. 91–95).

The difficulty of explaining extensive ponding on the sloping constructional surface of the High Plains has been pointed out by Smith (1940, pp. 90–91) and in itself could be considered fatal to the hypothesis of extensive lakes. There remains, however, the question of whether algal structures are present and, if so, of their distribution and relative importance. Because most of the features noted above either are definitely not of algal origin or are more readily explained by other processes, it is believed that, if algae were present at all, they were of very minor importance. Structures similar to the described "algal" structures have been found in the caliche of several successive physiographic surfaces. They are not a stratigraphic marker in southeastern New Mexico. There is no reason for considering the Llano caprock as distinct from caliche caprocks on younger surfaces. It differs only in exhibiting a more advanced stage of development.

Fluvial hypothesis.—Deposition of calcareous crusts along streams is known to occur in southern Texas and New Mexico and has been considered in combination with other processes (Trowbridge, 1926; Price, Elias, and Frye, 1946, p. 1746). Recent rimstone tufa and calcareous crusts on stream cobbles were observed by the writers in the Pecos River near Loving, New Mexico, and in Black River near Black River village, New Mexico (fig. 1). Encrustations of this type, however, have none of the features of the caliche profile and are distinct from the deposits described above.

Ascending ground-water hypotheses.—Deposition by capillary rise of water from the water table cannot be dismissed because of present low water tables in the area. These levels were much higher before dissection of the surfaces occurred and also during the pluvial stages of the Pleistocene. The facts which debar accepting this explanation include (1) solution-cupping of the upper surfaces of limestone pebbles; (2) occurrence of siliceous residuals; (3) downward penetration of caliche into impervious shales; and (4) occurrence of caliche on slopes where solutions could never be brought to the surface by capillarity. Under favorable physiographic conditions, capillary water may have augmented deposition from descending soil water in the lower part of the profile, but it could hardly explain the important lowering of the surface by weathering and solution which is indicated by the siliceous residuals.

Pedologic hypothesis.—Most of the features described in the present paper clearly indicate that soil-forming processes were of major importance in producing the caliche in southeastern New Mexico. The first five features noted at the beginning of this section strongly support this explanation. In particular, the caliche profile is a significant soil feature and very difficult to explain by any other hypothesis.

The profile development is believed to be best explained by conditions of alternating saturation and desiccation in a relatively dry climate with periodic rains. Solution occurred as the soil waters moved downward during a rainy period, and precipitation took place with the capillary rise of saturated alkaline solutions as the soil zone dried out. The solution-cupping on the upper surfaces of pebbles in the profile and caliche crusts on their undersurfaces clearly record these processes. The depth of the profile was probably determined by the maximum penetration of soil water and by the cumulative effect of periods of alternating saturation and desiccation of varying intensities. Variation in the depth and maturity of the profile on surfaces of different ages constitute a graded series and (fig. 3) appear to be due to these differences in age rather than to successively shorter independent periods of calichification. From this it follows that caliche formation continued after dissection of the higher surfaces and consequent regional lowering of the water table had occurred and therefore cannot be related to capillary rise of moisture from the water table. Overlapping of zones of solution and deposition is indicated by solution-cupped nodules in the upper part of the chalky caliche zone. A general lowering of the ground surface during caliche formation, as emphasized by Price (1933, p. 505), is evidenced by the siliceous residuals in the caliche, especially where the parent-limestone gravels have been completely removed by the process.

The banding, the brecciation, and the manganese oxide rims in the caprock are believed to record repeated cycles of desiccation and rainfall, with many light rains which penetrated only the caprock zone of the profile. Under these conditions brecciation could have resulted both from drying-out of the soil zone and from expansion accompanying impregnation of caliche. The manganese oxide rims could have formed as a desert varnish. It is probable that these processes went on under a thin soil cover, although at present the *A* zone has been stripped away and the *B* zone, as caliche, is exposed at the surface over wide areas.

AGE AND CLIMATIC IMPLICATIONS

Differences in the age of the caliche are evidenced by stratigraphic relations, physiographic position, and variations in the caliche profile. The Ogallala caprock in western Texas, which is continuous with that in the area studied in New Mexico, is older than the Blanco beds and is thus post–middle Pliocene to pre–early Pleistocene in age (Evans, 1948). Caliche on the Sacramento plain occupies a probably equivalent physiographic position and would be of comparable age. Younger caliches on the lower Mescalero–Diamond A, the Blackdom, and the Orchard Park plains cannot be dated so accurately but doubtless were initiated during interglacials of the Pleistocene. It is probable that profile formation, interrupted by climatic changes, has continued on all surfaces into Recent time.

The climatic implications of caliche as a polygenetic fossil soil have been pointed out by Bryan (1943), who considers successive caliches the record of relatively arid conditions, and solution features as the record of relatively humid conditions. This interpretation is supported by the occurrence of caliche in arid and semiarid regions throughout the world and seems much more probable than Theis's (1936) suggestion that the Ogallala caprock was deposited under Pleistocene pluvial conditions.

TABLE 2

TENTATIVE SEQUENCE OF PEDOLOGIC AND GEOLOGIC EVENTS

Time	Climate	Pedologic Events	Geologic Events
Recent	Relatively arid	Deposition of small caliche nodules in B zone of present soils	
	Relatively humid	Development of solution channels and degraded caliche	
	Relatively arid	Calichification of older solution-channel fills	
	Relatively humid	Development of older solution features	
Pleistocene	Alternating relatively humid and arid	Development of caliche on younger surfaces and continued development of Ogallala caprock	Erosion of Orchard Park, Blackdom, and Mescalero–Diamond A plains; deposition of Gatuna formation and Blanco beds
Late Pliocene	Relatively arid	Initial development of Ogallala caprock	
Middle and early Pliocene	Relatively humid		Deposition of Ogallala formation

On the basis of the above, the caliche sections in southeastern New Mexico indicate the sequence of events shown in table 2. This sequence is obviously tentative, and a definite correlation with the important climatic changes of the Pleistocene is unwarranted, largely because of uncertainties in dating erosion surfaces. The sequence is significant, however, in indicating that further investigations of complex caliche profiles, together with geomorphic studies, may reveal a relatively complete record of climatic changes in the region. The record provided by most caliche sections is likely to be incomplete, and it remains to discover critical sections and to correlate events over wider areas before a satisfactory chronology can be proposed.

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TOPOGRAPHY AND SEDIMENTS OF THE ARCTIC BASIN[1]

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ABSTRACT

Contours of the Arctic Basin were based on a compilation of all known soundings, largely from recent Russian expeditions. Additional general inferences regarding the bottom topography were derived from indirect considerations, such as distribution of Pleistocene glaciers, character of land topography, positions of earthquake epicenters, and sources and composition of the bottom sediments. It is concluded that the floor of the basin is far more irregular than is indicated by the present sparse soundings. Off Alaska and Canada the basin is deepest and least well sounded.

INTRODUCTION

The current international interest in the Arctic demands a more complete knowledge of that area than is generally available at present. One of the least-known aspects is the submarine topography of the Arctic Basin. To some extent it is possible to supplement the sparse soundings by indirect lines of evidence. The writer has attempted to summarize the existing information mainly in an effort to stimulate the collection of additional data. In this he has been aided by conversations with Dr. R. H. Fleming, of the United States Navy Hydrographic Office, and with Dr. R. S. Dietz, of the United States Navy Electronics Laboratory. Much general information about the Arctic is presented in a symposium on polar research edited by Joerg (1928), in a summary of Russian activity by Taracouzio (1938), and in a United States Navy orientation pamphlet by Shelesnyak (1947).

SOUNDINGS

The most direct method of determining the character of the sea-floor topography is through the study of soundings; but, in the Arctic, sounding operations have been hampered by sea ice. The edge of the permanent ice pack, as shown by the *Ice Atlas of the Northern Hemisphere*, coincides fairly well with the outer edge of the continental shelf. Few ships are strong enough to penetrate this ice, so soundings in the main basin are sparse. The continental shelf, however, is mostly covered only by seasonal ice; therefore, it is relatively open to shipping during the summer and is fairly well sounded. Because of their relative abundance, soundings shallower than 500 fathoms are not indicated on the accompanying chart, although the 100-fathom contour based on them is shown.

Ships which reach the edge of the permanent ice pack usually find depths suggestive of the beginning of the continental slope. In 1913 the "Karluk" took nine deep soundings in open water north of Alaska before becoming imprisoned in the ice and eventually being crushed off Siberia (Stefansson, 1921). While the "Karluk" was in the ice, several deep-sounding casts were made, but all failed to reach bottom. Similarly, a large number of additional soundings have been taken in open marginal waters near Spitsbergen, mostly by Danish ships.

The thickness of the permanent ice pack ranges up to about 15 feet, making it practically impenetrable to shipping.

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Consequently, soundings in the central area of deep water have had to be obtained by unusual methods of traveling on, above, below, or with the ice rather than through it. The earliest method of traveling in the Arctic was by sledge and dog team, and some soundings have been obtained on such trips. During his dash for the North Pole in 1909, Peary took several soundings north of Grant Land. He attempted a sounding within a few miles of the pole, but the line broke when 1,500 fathoms had been paid out. Stefansson (1921) in his various trips over the ice also made a number of soundings. Most of them, however, were in depths shallower than 500 fathoms and thus are not indicated on figure 1. His work was largely confined to the Beaufort Sea and the area west of Grant Land.

Another method of traversing the Arctic is that of freezing a ship into the ice so that it drifts with the ice pack. A number of ships not specifically built for such purposes have been accidentally imprisoned and finally crushed; but the drift of their wreckage and other considerations showed that the ice itself is moving toward the Greenland Sea, where it melts. To make use of this drift, the "Fram" under Nansen's leadership was intentionally frozen into the ice pack north of the New Siberian Islands in 1893. During its three-year drift toward the Greenland Sea, eight wire-sounding casts reached bottom. Similarly, the Soviet icebreaker, "G. Sedov," under R. L. Samoilovich, drifted from 1937 to 1939, while the "Sadko," under Ioffe, was in the ice during two periods, 1935 and 1937–1938. About 55 wire soundings accompanied dredgings and other oceanographic observations (Buynitsky, 1940; Zubov, 1940; Gorbunov, 1946). Three main areas were investigated: north of the New Siberian Islands, north of Fridtjof Nansen Land, and off the Taimir Peninsula.

A third method of sounding in the area of ice pack was initiated in 1927, when Wilkins landed an airplane on the ice about 550 miles northwest of Point Barrow. He obtained a single echo sounding of 2,975 fathoms, the deepest reported in the Arctic Basin. In 1941 four more echo soundings north of Wrangell Island were made, along with other observations, by Akkuratov, who landed on the ice at three different localities (Sverdrup, Johnson, and Fleming, 1942; Akkuratov, 1948).

A fourth, intermediate, method was followed in 1937 and 1938, when four Russian scientists under Papanin's leadership were landed near the North Pole and were supplied by airplanes as they drifted on the ice toward the Greenland Sea. Many observations, including 16 deep soundings, were made (Shirshov, 1938, 1940; Papanin, 1939). Papanin's book is of considerable interest because it shows how thoroughly even excellent scientific work in Russia may be pervaded by communistic ideology.

Still another method of sounding was proposed for the "Nautilus" expedition of 1931, whereby a submarine would travel beneath the ice, coming to the surface in the open leads of the ice pack.

Soundings obtained by these various methods comprise a total of 152 deeper than 500 fathoms. This number amounts to only one per 12,000 square statute miles of area; but, unfortunately, instead of being well distributed, most of the soundings are concentrated in a few small areas. The positions of all soundings deeper than 500 fathoms are indicated by small circles on figure 1. Because of the difficulties of navigation in the Arctic, some of these positions are likely to be of rather low accuracy. The wire-sounding casts which failed to reach bottom are not shown, although they are of some value as indicators of minimum depth.

Depth contours at 500-fathom intervals were drawn for the Arctic Basin and for comparison in part of the north Atlantic Ocean as well. The 100-fathom contour where known is given as a dashed line. Off northeast Greenland it was taken from a new chart by Kiilerich (1945), who apparently based his contours on some unpublished Danish soundings. It is notable that the general form of the basin, as indicated by present data, is little different from that inferred by Nansen (1904) on the basis of only a few soundings. Where soundings are sparse, the present contours are smoothly curving, but in densely sounded areas they are more irregular and probably more nearly depict the true bottom configuration. The soundings taken by Akkuratov about 500 miles north of Wrangell Island are of especial interest, because they indicate the presence of a deep ridge which may extend completely across the Arctic Basin. The deepest part of the basin is near Bering Strait, in the least well-sounded region. Two recent contour maps show irregular and complicated topography in this area because the cartographers failed to recognize that the small bar and dot printed above some of the soundings on various charts indicate that no bottom was reached within the length of the available sounding wire. A 3,000-fathom contour was drawn in the bottom of the basin on the basis of the near-by deepest sounding of 2,975 fathoms.

GLACIAL EROSION

During the Pleistocene epoch glaciers from Greenland, Canada, and parts of Siberia radiated outward from their growth centers. Movement north of the centers was toward the Arctic Ocean, as shown by radiating glacial striae and other markings. According to Flint (1945, 1947, pl. 3), they reached beyond the present coasts from the Taimir Peninsula westward past Norway and Greenland to eastern Alaska. A minor exception is the northernmost tip of Greenland, which is unglaciated. Along most of this distance typical fiord shorelines are common. Off-shore islands in this area, such as North Land, Nova Zemlya, Fridtjof Nansen Land, and Spitsbergen were also glaciated. It is noteworthy that the sea floor between the islands and the mainland is very irregular and contains numerous troughs, which Nansen (1904) long ago recognized as being of probably glacial origin. The heads of some troughs extend close to the fiord coasts of the land and thus may be extensions of the fiords. Scattered about between the troughs are many banks and small islands, which may well be erosional remnants of glaciation. Though the area is not densely sounded, it is evident from the available data that this part of the shelf is much more irregular than are normal continental shelves. The largest areas of irregular shelf are in the Barents and the Kara seas, which are unusually deep shelves, generally exceeding 150 fathoms. Because of these relationships, figure 1 indicates that the areas covered by the ice sheet should possibly be extended beyond the limits shown by Flint's maps to include both the Barents Sea and the Kara Sea. The erosion could have taken place in water shallower than at present if it had occurred before isostatic depression of the shelf beneath the ice load. The glaciers in part may have moved northward from the mainland, as is suggested by the long trough east of Nova Zemlya. The fiord coasts of the islands indicate supplementary glacial erosion from these centers.

The shelf off Greenland and Canada is much less well sounded, but it is apparent that at least some of the deep troughs which separate the land into large islands extend seaward across the narrow shelf. Spencer in 1905 ascribed some of these submarine troughs in part to glacial erosion. Since that time almost no additional soundings have been obtained to show whether other troughs are present.

From eastern Alaska westward past Bering Strait to Taimir Peninsula the continental shelf is shallower, mostly less than 80 fathoms, and extremely flat. This flatness was noted by Nansen, who remarked that ships sailing off Siberia were subject to little danger of running aground, even though the water is not deep. Recent investigations with recording echo-sounders in the Bering Sea and the Chukchi Sea indicate that these areas are among the flattest in the world (LaFond, Dietz, and Prichard, 1949). This flatness, together with the fact that the northern border of the Pleistocene ice sheet in both Siberia and Alaska is well inland, suggests that the continental shelf there was never subjected to much glacial erosion, although Flint indicates that Wrangell Island and the New Siberian Islands were covered by glaciers.

TOPOGRAPHY OF ADJOINING LAND

Some information regarding sea-floor topography may be provided by the seaward projection of land mountain ranges. A general topographic map of the North Polar region by Joerg (1930) shows that the Brooks Range of Alaska and the Anadyr Mountains of Siberia are essentially parallel to the coast and do not extend seaward. This is the usual relationship for Suess's Pacific type of coasts. The Khayakhak and Kharaulakh Mountains of Siberia and the Ural Mountains, which continue through Nova Zemlya, however, are almost at right angles to the coast (Atlantic type) and may therefore be reflected in local outward projections of the Arctic Basin margins, like the one north of the New Siberian Islands. The various ranges are separated by geologically quiescent shield areas, from which one would expect few structural projections into the basin. These shield areas are also characterized by broad, nearly flat plains bordering the ocean. An excellent discussion of the general geology of the Arctic region was presented by Eardley (1948), who drew on structure, stratigraphy, and paleontology to outline the paleogeography of the area. He concluded that the basin is underlain by continental rocks which began to sink in late Paleozoic time. Gutenberg and Richter (1939) state, however, that the amplitudes of reflected longitudinal earthquake waves probably indicate the presence of bottom material like that of the Pacific Ocean and unlike that of the continents. Unfortunately, the seismic velocity measurements needed to verify the character of the basement are lacking at present (Gutenberg and Richter, personal communication).

EARTHQUAKE EPICENTERS

It is well known that most earthquake epicenters fall along fairly well-defined belts, which also are areas of volcanic activity and mountain-building and thus are characterized by irregular rugged topography. The Mid-Atlantic Ridge is such an area. Various Russian workers and others have shown that the belt of epicenters related to the Mid-Atlantic Ridge continues northward from south of Iceland past Spitsbergen and along the European side of the Arctic Basin to the Lena River region, where it may meet the Kharaulakh and the Khayakhak mountain ranges. Figure 1 shows the epicenter belt taken from a map reproduced by Heck (1938) from Russian sources. This is based on seismic data prior to about 1925. Gutenberg and Richter (1941, pp. 70–73; 1945) indicate that some of these epicenters may be poorly located, but the epicenters which they do consider acceptable follow the same general trend, except for a few scattered ones, chiefly near Bering Strait. The latter belong to the circum-Pacific belt of activity. Their epicenters, shown by large dots on figure 1, are of earthquakes which occurred between 1921 and 1944. Because most of them postdate the belt shown by Heck, they serve as an excellent check. All but two of the reported shocks are of less than magnitude 7, and consequently they are not considered major earthquakes. The fact that the belt extends from an irregular submarine mountain range across the poorly known Arctic Basin suggests that the topography of the part of the basin through which the belt passes may also be irregular.

SEDIMENTS

It appears that sediment descriptions for only six samples from deeper than 500 fathoms in the Arctic Basin are available; although it is possible that data for others have been published in Russian reports inaccessible at present. Five of the samples were collected aboard the "Fram" in 1893–1895 from northwest of the New Siberian Islands. Four of these are from depths between 1,950 and 2,200 fathoms. The sediments are very fine-grained, containing less than 0.1 per cent of grains coarser than 1 mm. and more than 60 per cent of grains finer than 0.01 mm. (Bøggild, 1906). All were named "gray clay," although a strong brown tinge was present. Only one sample contained appreciable organic material, but even this one had less than 4.5 per cent of *Globigerina* tests. The description of grain size, grain composition, and chemical analysis is such that all five samples could easily be "red clay," like the "red clay" of the Pacific Ocean, which also is more brown than red. One of the deep samples which was well described is summarized in table 1, with values recomputed from the original for greater ease of comparison. It should be noted that the chemical composition is just about the same as that given by Sverdrup, Johnson, and Fleming (1942) for 51 samples of "red clay" from the Atlantic, Pacific, and Indian oceans (Steiger's analyses). This particular sample, however, contains nearly equal amounts of ferric and ferrous iron, unlike most of the other Arctic samples, for which ferric is far in excess of ferrous iron—as is characteristic of "red clay." The mineral composition is within the common range of variation of "red clay." Most striking is the absence of volcanic glass shards and of manganese minerals, although the latter may possibly be included with opaque or aggregate minerals. Calcite was present as foraminiferal tests, almost exclusively of *Globigerina*. According to Brontman (1938, p. 277), a sample taken by the Russian North Polar Expedition at lat. 87°10′ N., long. 1° E., 2,381 fathoms, consisted of silt, reddish-brown at the top and gray below. Wüst (1942) also reported that, of 39 deep samples obtained by the "Sedov," at least 26 were brown.

Ten more sediment samples were collected aboard the "Fram" from the continental shelf between the Kara Sea and the New Siberian Islands. All are fine-grained, some as fine as the samples from deep water. Most are also characterized by a brownish color. Nineteen cores obtained by the "Belgica" from the shelf near Nova Zemlya consist of brown sediment (Thoulet, 1910). Four additional samples from the shelf of the Chukchi Sea were taken by the United States Navy Electronics Laboratory in 1947 (LaFond, Dietz, and Prichard, 1949). Mechanical analyses showed that three of them are very fine-grained, having median diameters between 0.0035 and 0.009 mm.; the fourth sample is much coarser. Limy and siliceous remains of organisms are minor constituents, usually making up less than 1 per cent of the sample. Trask (1932) described 19 more samples from shallow near-shore waters along Alaska and in the Canadian Archipelago and 29 others from deeper water near the south end of Baffin Bay. Nearly all are gray, and many are poorly sorted. Several hundred additional samples from the Barents Sea and the White Sea were briefly described by Gorshkova (1931). Many of these are poorly sorted, and some are brown-colored and contain a high percentage of manganese oxide.

TABLE 1

COMPOSITION OF SEDIMENT FROM LAT. 83°24′ N.; LONG. 102°14′ E.; 1,900 FATHOMS

MINERAL GRAINS COARSER THAN 0.05 MM.	Per Cent	CHEMICAL COMPOSITION		
		Oxide	Arctic Sample (Per Cent)	Average of 51 "Red Clay" Samples (Per Cent)
Calcite	3.83	SiO_2	57.21	54.48
Quartz	2.45	Al_2O_3	15.64	15.94
Augite	0.92	Fe_2O_3	5.05	8.66
Aggregates	0.79	FeO	4.25	0.84
Opaque	0.50	MnO	0.58	......
Hornblende	0.32	MnO_2	0.05	1.21
Orthoclase	0.23	CaO	3.56	1. 6
Microcline	0 20	MgO	1.69	3.31
Olivine	0.15	K_2O	2.7	2.85
Plagioclase	0.10	Na_2O	1.96	2.05
Garnet	0.03	P_2O_5	0.26	0.30
Epidote	0.02	H_2O	5.03	7.04
Tourmaline	0.02	Others	2.20	1.36
Zircon	0.02			
Total	9.58	Total	100.22	100.00

A large amount of additional information on bottom materials of the shelf is available in the form of bottom notations on the excellent large-scale navigational charts recently published by the United States Navy Hydrographic Office (HO 6601-6). These notations are based on the small quantity of sediment which adheres to the tallowed lead weight on sounding lines. Because many different observers collected the data, such notations must be used with caution. Altogether, 158 bottom notations were counted; of these, 115 are listed as "mud" or "clay." Of the 47 notations for which the color is indicated, 32 are brown. Most of these chart notations of color were based on brief sample studies by Bissett (1930), who found that, in the reddish or brownish mud, FeO and Fe_2O_3 constitute 0.13 and 8.14 per cent, respectively, while in the blue or gray mud the values are 3.22 and 0.89 per cent. It is clear, therefore, that the brown color is due to oxidation of the iron. The other bottom notations on the charts are as follows: mud-and-sand, 9; sand, 11; stony, 4; rocky, 8; and various mixtures of mud with stony and rocky, 9. It is notable that the most homogeneous bottom is that of the shelf off Siberia. This area is characterized almost exclusively by mud, except near islands.

There are several agents and corresponding source areas which may contribute sediments to the Arctic Basin. One of these agents is ice. Available evidence indicates that continental glaciers did not reach the present coastline from eastern Alaska westward to the Taimir Peninsula; thus this area of continental shelf and basin could have received no glacial till. In the rest of the circumference of the Arctic Ocean, glaciers did reach the present coast. Judging from the topography of the shelf of the Barents and Kara seas, the glaciers did much active excavation there. Most of the chart notations of mud mixed with stony or rocky material occur in these areas. The poor sorting is suggestive of till, perhaps deposited atop the eroded surface by the retreating glacier. Much more morainic material must have been deposited beyond the zones of excavation, possibly at the edge of the shelf.

Ice-rafting of sediments is widely reported from the Beaufort Sea by Kindle (1924); off Greenland by Tarr (1897) and others; in the East Siberian Sea by Sverdrup (1931); and from the Chukchi Sea by LaFond, Dietz, and Prichard (1949). The sediments are carried primarily by ice floes, which pick them up by bottom-freezing or by ice-shove against the shore. Some sediment transportation is also accomplished by icebergs from glaciers which reach the sea and calve off. River ice is also important. It is noteworthy that these reports of sediment-laden ice are from marginal areas. Nansen (1904) remarked particularly on the absence of coarse sediments on the ice above the deep water of the basin but stated that embedded mud was present. In his opinion, most of this material is transported completely across the basin and deposited east of Greenland, where the ice melts. He considered that only a small amount of it can be released above the basin by bottom-melting or by overturning of the ice.

During the short summer season the melting ice of the land brings the streams to flood, and then they transport much sediment to the coast. The glaciated half of the Arctic coast contains many fiords and deep troughs which serve as settling areas; thus practically no stream-borne sediment should reach the shelf off Europe, Greenland, and northeastern Canada. The rest of the coast, however, is characterized by a number of large streams, such as the Mackenzie and the Lena rivers, which are actively building deltas on the shelf. As these rivers are of low gradient, they carry much fine-grained material, which may reach well beyond the deltas. Possibly this river-supplied fine material is responsible for the wide expanse of mud bottom of the East Siberian Sea. Only a minor contribution is made by wave erosion because waves and even tides are of very small height along most of the shore, owing to friction losses in crossing the wide shelf and to damping action by the ice.

A number of authors have remarked on the strength of the wind radiating outward from the Greenland icecap (Kindle, 1924; Matthes, 1946; Hobbs, 1948), though there is disagreement regarding its origin. Mention is made of the transportation by wind of coarse debris onto the ice to a distance of several miles from land. Thin deposits of dust on the ice many miles from shore have also been noted by Kindle and others, who believed them to consist of volcanic ash or perhaps even of cosmic dust. Prior to present wind transport, widespread deposits of loess were formed during the Pleistocene epoch, when the glaciers were present. Hobbs showed that a wide belt of loess north of the glaciers in Siberia was deposited by northward-blowing winds. The site of similar deposits from Taimir Peninsula westward to eastern Alaska would now be below sea level on the shelf or in the Arctic Basin.

A minor contribution to the sediments of the Arctic is made by inflowing Gulf Stream water carrying various warmer-water organisms. Sponge spicules and other organic debris found in sediments just north of Europe are believed to have this origin (Veen, 1913). The Gulf Stream water continues on into the Arctic Basin, where it forms a layer of relatively warm water sandwiched between top and bottom waters having temperatures below 0° C.

In contrast to the north Atlantic and north Pacific oceans, the deep polar sea appears to be very deficient both in species and in numbers of planktonic algae (Gran, 1900). Nansen considered this lack the result of light reflection and absorption by the thick ice cover. Some discoloration of the ice by diatoms was believed to have resulted from attachment or freezing of the organisms into the ice nearer shore. In the absence of plants, there should be few animals, a conclusion which is confirmed by Bogarov (1946), who reported that study of the "Sedov" material showed that scanty zoöplankton is characteristic of the central part of the Arctic Ocean. This explains the very low percentage of foraminiferal tests in the sediments, although these samples came from depths at which *Globigerina* ooze occurs elsewhere. The discovery of a few polar bears and seals on the ice pack (Stefansson, 1921; Akkuratov, 1948; and others) appears to be contradictory evidence but may mean simply that where the ice cover is locally absent, as in the open leads, life can be abundant. The low percentage of calcium carbonate in the sediments is in agreement with Trask's (1937) statistical studies for other oceans, which indicated that, where surface-water salinities are lower than 34 parts per thousand, the calcium carbonate of the sediments is generally less than 5 per cent. Shirshov (1940) showed that the surface salinities in the Arctic are less than 31.5 parts per thousand.

In summary, a number of agents deposit sediments in the Arctic Ocean. The sea floor between Taimir Peninsula westward to eastern Alaska should be characterized by former direct ice deposits and former wind-laid deposits. The remainder of the coast now receives most of the stream-borne sediments. Ice-rafted sediment and organic debris may be evenly distributed between the two areas. In the writer's opinion, this distribution of effective agents probably is reflected in a somewhat greater rate of deposition in the basin nearest northern Europe than elsewhere. It is noteworthy in this connection that the floor of the basin appears to be somewhat shallower there than off Alaska and eastern Siberia. The fine-grained texture of most of the sediments, both in the basin and on the shelf, together with the prevalent brownish oxidized color, indicates that deposition is very slow, at least at the present time. Probably deposition takes place so slowly that irregularities formed by faulting have not been buried, and therefore the bottom of the basin, particularly in the belt of epicenters, is far more irregular than is indicated by the generalized contours of figure 1.

CONCLUSIONS

The following tentative conclusions regarding the topography of the Arctic Basin are based on the scanty data outlined above.

1. The shelf and margin of the basin from the Taimir Peninsula westward to eastern Alaska is deep and irregular because of glacial erosion and deposition, while the rest of the shelf is shallow and flat because it was not covered by glaciers.
2. The slope and floor of the basin off Europe may be irregular because of submarine faulting.
3. The basin is oval-shaped, perhaps deepest near Alaska and eastern Siberia, and it may be divided by a structural ridge extending northward from a position between Wrangell Island and the New Siberian Islands.
4. The floor of the basin may be fairly irregular because of the present slow rate of deposition of sediments by all agents.

Unquestionably, much more should be learned of the basin. High on the list stands the need for additional soundings. A line of soundings across the least-known area could be obtained by freezing a ship into the ice near the Beaufort Sea. As pointed out by Zubov (1940) and others, a ship starting there would pass very near the Pole. An excessive amount of time, however, would be required for such a drift. Possibly the soundings can best be obtained by landing small airplanes carrying portable sonic sounders onto the ice at a number of points. This method might be most adaptable to the large unsounded area off Alaska, where airplane facilities have already been established. A method which would yield continuous profiles rather than separate spot soundings would be sounding from a submarine traveling completely submerged beneath the ice, a plan which was contemplated for the "Nautilus" expedition in 1931. Regardless of how they are obtained, future soundings and bottom samples will be valuable to geological knowledge of the Arctic.

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REPLACEMENT AND THE EQUATING OF VOLUME AND WEIGHT[1]

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ABSTRACT

A method is presented for writing molecularly, volumetrically, and electrically balanced equations for replacement reactions. Two general situations obtain: (*A*) enough is known of the chemistry of the particular reaction to write a complete equation, or (*B*) only enough is known to permit the setting-down of the essential facts of the reaction in skeleton form. In the first case a simple, molecularly balanced equation for the reaction in question will show that there are three different possibilities: (1) that too much guest is furnished for the volume left by the departing host; (2) that too little guest is produced to fill the volume of host removed; and (3) that the volumes involved are approximately equal. In the second case the balanced skeleton equation will show that (1) there is a deficiency in the host of the large ions (sulfur, oxygen) needed in an equal volume of the guest; (2) there is an excess of these large ions in the host over the amount that can be used in an equal amount of guest; and (3) the number of large ions in equal volumes of host and guest are the same. In both cases the third category is relatively rare. Methods by which the additional material needed can be provided or the extra material can be removed are suggested. In supergene sulfide replacements the main problems in achieving complete balance are (*a*) a mechanism for the oxidation of excess S^{-2}, (*b*) a source of hydrogen sulfide, and (*c*) a mechanism for the reduction of S^{+6} in solution to provide additional S^{-2}. A naturally present reducing agent which can do the required work is not immediately apparent; the most likely oxidizing agent is Fe^{+3}; H_2S probably derives from the action of H_2SO_4 on certain sulfides. Hypogene sulfide and sulfosalt metasomatic reactions cannot be written in complete form except in a few cases. If such replacements require that sulfur be removed, it is considered to be oxidized to the sulfate ion. If additional sulfur is needed, it is provided directly from the ore-bearing solutions. No attempt is made to show how it got to the scene of deposition. Other replacements than those of sulfides and sulfosalts are produced in equational form, and similar conclusions are reached as to the requirements for putting them in complete balance.

INTRODUCTION

In a recent paper Buerger (1948, p. 114) remarked that "diffusion suggests the mechanism" for volume-for-volume replacement. He goes on to say that replacement is "substantially a matter of diffusion of new metals into the volumes dominated by oxygen or sulfur atoms." This statement is true enough when oxides, silicates, or carbonates are replacing silicates or when sulfides or sulfosalts are replacing sulfides. But it is not applicable to the replacement of silicates, oxides, or carbonates by sulfides. When such replacements occur, sulfur must be substituted for oxygen, and this is a far more drastic process than the exchange of metal ions which are far smaller and whose movements have a smaller effect on the structure of the host mineral. But, because sulfides do replace other classes of minerals, this sulfur-for-oxygen substitution must take place, comparatively difficult though it may be to carry out. Once the guest ion or ions have been brought into contact with the host mineral, it seems reasonable that the actual replacement is accomplished through new material being diffused into, and old material out of, the space originally occupied by the original mineral.

THE PROBLEM OF VOLUME-FOR-VOLUME REPLACEMENT

The acceptance of Buerger's general principle does not solve one problem which has been puzzling geologists ever since it was first realized that replacement was essentially a volume-for-volume exchange. That problem is to write chemical equations for replacement reactions which, although representing possible and molecularly balanced reactions, also contain equal volumes of solid material on both sides of the equation. The present concept, if the textbooks are consulted, is to say that the equation given indicates what happens in a general way during metasomatism but that, because the phenomenon occurs on a volume-for-volume basis, it does not give an exact representation of the chemistry of the process. A simple example of such an equation would be

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[1] Contribution of the Division of Geology, Pennsylvania State College. Manuscript received November 9, 1948.

$$2Ag^{+1} + SO_4^{-2} + CuS = Ag_2S + Cu^{+2} + SO_4^{-2}\,. \qquad (1)$$

The actual space occupied by 1 formula unit[2] (molecule) of covellite is 34.24 cubic Ångstrom units (see table 1), whereas that taken up by 1 molecule of argentite is 58.11 Å^3. If the replacement of covellite by argentite occurred exactly as represented by equation (1), there would be a gain of very nearly 70 per cent in volume during the process. Because such a gain of volume does not take place under the conditions of hypogene or supergene metasomatism, it follows that equation (1) must be modified so as to equate not only molecular units but unit volumes as well. In other words, a quantitative approach to the problem should be attempted, in contrast to the qualitative one of equation (1).

SULFIDE AND SULFOSALT REPLACEMENTS—SUPERGENE CONDITIONS

GUEST IN EXCESS

In the preceding paragraph it was shown that the replacement of 1 molecule of covellite (Vol. = 34.24 Å^3) by 1 molecule of argentite (Vol. = 58.11 Å^3) would find 23.87 Å^3 of argentite for which no space would be provided by the removal of 1 formula unit of covellite. These 23.87 Å^3 are 70 per cent of 1 formula unit volume of covellite. This means that 1.7 molecules of covellite must be removed for each molecule of argentite formed by the metasomatic process if no loss of volume is to occur. To eliminate fractions, it would be necessary to multiply the 1.7 molecules of covellite and the 1 molecule of argentite by 10; i.e., 17 molecules of covellite must be taken away for each 10 molecules of argentite deposited. But how are these extra molecules of covellite to be removed? Because the probable solubility of cupric sulfide in the solutions accomplishing this supergene replacement is vanishingly small, it is unlikely that the excess covellite can be removed in simple solution. Some other mechanism, therefore, must be suggested. It would seem that this could easily be the oxidation of the S^{-2} to the S^{+6} of the sulfate radical. As the necessary concomitant of this oxidation, the reduction of ferric ion to ferrous ion should be adequate, as is explained

[2] The unit cell of the average mineral contains more than one molecule. If the volume of the unit cell is divided by the number of molecules (formula units) contained in it, the result will be the volume, in cubic Ångstrom units, of one formula unit of the mineral in question. Fairbairn (1943, pp. 1308–1310) gives a brief and clear explanation of the calculation of unit cell volumes and of the relation of these volumes to density. The formula for the unit cell volume of a rhombohedral mineral is given incorrectly, however, and should be $a^3\sqrt{1 - 3\cos^2 a + 2\cos^3 a}$. In table 1 are given the unit cell volume, the number of formula units in the unit cell, and the volume of the formula unit for each mineral with which this paper deals. The data for calculating these values have been obtained for sulfides, oxides, and native metals from Dana (1944). For calcite, fluorite, alpha quartz, and wollastonite, the necessary data have been taken from Bragg (1937). The results have been compared, when possible, with Fairbairn's (1943, pp. 1317–1319) and are in close agreement.

TABLE 1

Mineral	Unit Cell Volume (Å³)	No. of Formula Units in Cell	Volume per Formula Unit (Å³)
Anglesite ($PbSO_4$)	313.92	4	78.48
Argentite (Ag_2S)	116.21	2	58.11
Bornite ($CuS \cdot 2Cu_2S \cdot FeS$)	1,305.75	8	163.21
Bournonite $\frac{1}{2}(2PbS \cdot Cu_2S \cdot Sb_2S_3)$	543.00	4	135.75
Calcite ($CaCO_3$)	122.06	2	61.03
Chalcopyrite ($CuS \cdot FeS$)	282.81	4	70.70
Chalcocite (Cu_2S)	7,285.79	160	45.54
Copper (Cu)	46.79	4	11.70
Covellite (CuS)	205.46	6	34.24
Enargite $\frac{1}{2}(Cu_2S \cdot 4CuS \cdot As_2S_3)$ or $(3Cu_2S \cdot As_2S_5)$	296.63	2	148.31
Fluorite (CaF_2)	161.82	4	40.46
Galena (PbS)	208.53	4	52.13
Jamesonite ($4PbS \cdot FeS \cdot 3Sb_2S_3$)	1,195.26	2	597.63
Magnetite ($FeO \cdot Fe_2O_3$)	587.22	8	73.40
Polybasite ($4Ag_2S \cdot 4Cu_2S \cdot Sb_2S_3$)	1,164.23	2	582.11
Pyrite ($FeS \cdot S$)	157.90	4	39.48
Quartz (alpha) (SiO_2)	112.28	3	37.43
Silver (Ag)	67.78	4	16.95
Sphalerite (ZnS)	157.46	4	39.37
Stannite ($Cu_2S \cdot FeS \cdot SnS_2$)	319.73	4	79.93
Tennantite ($5Cu_2S \cdot 2CuS \cdot 2As_2S_3$)	1,058.09	2	529.04
Tetrahedrite ($5Cu_2S \cdot 2CuS \cdot 2Sb_2S_3$)	1,102.21	2	551.10
Wollastonite ($CaSiO_3$)	390.00	6	65.00
Zincite (ZnO)	47.13	2	23.56

shortly. To provide the needed ferric ion, it seems reasonable to assume that some ferric sulfate or chloride may have survived the trip down through the zone of oxidation to react with the excess of covellite to form sulfate ion from the sulfide and to be itself changed to ferrous iron. This can be expressed in an equation as follows:[3]

$$\left.\begin{array}{l} \mathrm{17CuS + 20Ag^{+1} + 10SO_4^{-2} + 56Fe^{+3} + 168Cl^{-1} + 28H_2O} \rightarrow \\ (582.08\ \text{Å}^3) \\ \qquad\qquad \mathrm{10Ag_2S + 17Cu^{+2} + 17SO_4^{-2} + 56Fe^{+2} + 112Cl^{-1} + 56H^{+1} + 56Cl^{-1}}, \\ \qquad\qquad (581.10\ \text{Å}^3) \end{array}\right\} (2a)$$

Volume difference = 0.17 per cent.[4]

The equation just given, (*2a*), can be expressed by far fewer molecules if it is considered acceptable to have a volume difference of between 1 and 2 per cent instead of the quite small amount resulting in equation (*2a*). This is illustrated by equation (*2b*):

$$\left.\begin{array}{l} \mathrm{5CuS + 6Ag^{+1} + 3SO_4^{-2} + 16Fe^{+3} + 24SO_4^{-2} + 8H_2O} \rightarrow \\ (171.20\ \text{Å}^3) \\ \qquad\qquad \mathrm{3Ag_2S + 5Cu^{+2} + 5SO_4^{-2} + 16Fe^{+2} + 16SO_4^{-2} + 16H^{+1} + 8SO_4^{-2}}, \\ \qquad\qquad (174.33\ \text{Å}^3) \end{array}\right\} (2b)$$

Volume difference = 1.80 per cent.

Here use has been made of ferric sulfate instead of ferric chloride, but the principle remains the same because the oxidizing agent in both instances is ferric iron.

These last two equations—(*2a*) and (*2b*)—depart from the accepted thought on the process of replacement, in that they introduce components on both sides of the equation which, while taking a very definite part in producing a volumetrically balanced product, leave no solid evidence of their presence behind them. This departure from the simple, molecularly balanced equation seems well grounded, in that it permits the equation to equate volumes as well as molecular weights and electrons. The attempt made in these two equations to explain the process by which the excess molecules of covellite, which must be removed to achieve volumetric balance, are got into solution is not backed by experimental proof, but that lack does not invalidate the general method of balancing replacement equations here presented. It is, nevertheless, advisable to consider the possibility of the occurrence of the reaction of equations (*2a*) and (*2b*).

In this paper most of the supergene equations considered are of the type of equation (*2a*). If the equations chosen are representative of supergene reactions, then it is more common for the volume of the replacing mineral or minerals produced (by a molecularly balanced equation) to be greater than that provided by the removal of the replaced mineral. To achieve volumetric balance in equations of this kind, solid sulfides, above the amount required for molecular balance of the basic equation, must be dissolved and removed in solution. Primarily, this is a problem of oxidizing some of the sulfide ion of the replaced mineral to the sexivalent sulfur of the sulfate radical. This, it has been assumed, is accomplished by the reduction of Fe^{+3} to Fe^{+2}, the emf of the cell so set up being sufficient to drive the various stages of the reaction forward. It is now advisable to consider in some detail the adequacy of the oxidation potentials of the electronic reactions suggested as taking place when a sulfide ion is converted to sulfate. The half-reactions or electronic reactions used in this section and in the other parts of this paper are given in table 2. From these it can be seen that the oxidation of sulfide ion to sulfate can go forward under the standard conditions there listed (effective unit concentrations, 25° C., and 1 atm. of pressure) when the other half of the oxidation-reduction pair is ferric to ferrous iron. The reaction proceeds in three stages, which are given herewith (for the sake of completeness, sphalerite is used as the initial source of sulfide ion):

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[3] It will be noted that, in this equation and those of the same type which follow, no attempt has been made to achieve balance in terms of the numbers of molecules in the unit cells of the minerals taking part in the reaction. Such unit cell equality can, however, be attained merely by multiplying through the equation by the largest number (of molecules) contained in any one of the unit cells involved. In eq. (*2a*) this number would be six.

[4] It will be found that this equation and those of the same type which follow it are not quite in volumetric balance. The deviations from equality usually are quite small (less than 2 parts in 100), but some have differences up to 3 per cent. Such differences are larger than the limits of error in the determinations of the dimensions of the unit cells used but are allowed to stand because exact volumetric balance would require the use of integers tens or hundreds of times larger than those here accepted.

TABLE 2*

Electronic (or Half-) Reaction	Oxidation Potential (Volts)†	Electronic (or Half-) Reaction	Oxidation Potential (Volts)†
$SO_3^{-2}+2OH^{-1}=SO_4^{-2}+H_2O+2e^{-1}$	0.90	$Mo^{+3}(r)=Mo^{+5}+2e^{-1}$	−0.11
$S+6OH^{-1}=SO_3^{-2}+3H_2O+3e^{-1}$	0.61	$H_2S\rightarrow S+2H^{+1}+2e^{-1}$	−0.14
$Cd_2^{+2}=2Cd^{+2}+2e^{-1}$	0.6	$Sn^{+2}\rightarrow Sn^{+4}+2e^{-1}$	−0.15
$Fe(OH)_2+OH^{-1}=Fe(OH)_3+e^{-1}$	0.56	$Cu^{+1}\rightarrow Cu^{+2}+e^{-1}$	−0.17
$S^{-2}=S+2e^{-1}$	0.51	$H_2O+H_2SO_3\rightarrow SO_4^{-2}+4H^{+1}+2e^{-1}$	−0.20
$HS^{-1}+OH^{-1}=S+H_2O+2e^{-1}$	0.48	$Cu\rightarrow Cu^{+2}+2e^{-1}$	−0.345
$Cr^{+2}=Cr^{+3}+e^{-1}$	0.41	$S+3H_2O\rightarrow H_2SO_3+4H^{+1}+4e^{-1}$	−0.45
$Ti^{+2}=Ti^{+3}+e^{-1}$	0.37	$Cu\rightarrow Cu^{+1}+e^{-1}$	−0.52
$Mo^{+3}(g)=Mo^{+5}+2e^{-1}$	0.25	$Fe^{+2}\rightarrow Fe^{+3}+e^{-1}$	−0.77
$V^{+2}=V^{+3}+e^{-1}$	0.20	$Ag\rightarrow Ag^{+1}+e^{-1}$	−0.80
$H_2=2H^{+1}+2e^{-1}$	0.00	$Mn^{+2}\rightarrow Mn^{+3}+e^{-1}$	−1.51

* The data contained in this table are taken from Latimer and Hildebrand, 1940, pp. 474–478; and Latimer, 1938, pp. 294–301.

† The oxidation potentials are for unit activities at 25° C. (298° K.) and 1 atm. of pressure.

$$ZnS + 2Fe^{+3} + 3SO_4^{-2} \rightarrow 2Fe^{+2} + Zn^{+2} + 3SO_4^{-2} + S\,, \tag{3a}$$

$$S + 4Fe^{+3} + 6SO_4^{-2} + 3H_2O \rightarrow 4Fe^{+2} + 4SO_4^{-2} + 6H^{+1} + 2SO_4^{-2} + SO_3^{-2}\,, \tag{3b}$$

$$2H^{+1} + SO_3^{-2} + 2Fe^{+3} + 3SO_4^{-2} + H_2O \rightarrow 2Fe^{+2} + 4H^{+1} + 4SO_4^{-2}\,. \tag{3c}$$

By adding algebraically the oxidation potentials of the two electronic reactions involved in each of these equations, the net emf in each can be determined. Under standard conditions in acid solutions, these emf are:

$$-0.14 - (-0.77) = +0.63 \text{ volt}\,, \tag{3a'}$$

$$-0.45 - (-0.77) = +0.32 \text{ volt}\,, \tag{3b'}$$

$$-0.20 - (-0.77) = +0.57 \text{ volt}\,. \tag{3c'}$$

Because each product is a positive voltage, the reaction will go in the direction indicated in each instance. This obviously does not prove that these reactions will go to the right under all conditions. In certain supergene solutions of low pH and relatively low iron concentration the reaction of equation (3*b*) at least might be reversed, but even this is unlikely. This discussion does show, however, that the reactions of the last three equations are possible and does nothing to demonstrate that they do not often occur.

In several of the equations to be taken up shortly, the oxidation of the sulfide to the sulfate ion is not accomplished by the reduction of ferric to ferrous iron alone. Rather, the process is carried forward in part by one or more of the following reductions acting in conjunction with that of ferric ion: cupric ion to cuprous, cupric to native copper, cuprous to copper, and silver ion to native silver. It is assumed that the emf of the various half-reactions remains the same in such mixed cells, unless that mixing also entails a change in effective concentrations. It is also believed that the tendency of one reaction to go in one direction may be more than counterbalanced by that of another to go in the opposite direction. For example, the emf of the electronic reaction of ferrous to ferric ion is +0.77 volt, whereas that of copper to cupric ion is −0.345 volt. As was shown by the calculations based on equation (3*b*), the reduction of ferric to ferrous iron can drive forward the oxidation of sulfur to sulfite ion. It is equally apparent from equations (3*d*) and (3*e*) and the calculations which follow them that the reduction of cupric ion to native copper does not produce a sufficient emf to drive the second and third stages of the sulfur oxidation forward. This is particularly apparent if it is noted from table 2 that the reduction of cupric ion to native copper actually is carried out in two stages—cupric to cuprous and cuprous to copper. The oxidation potential of the half-reaction converting cuprous to cupric ion is −0.17 volt, which is higher than the potentials for either the second or the third stage of the sulfur oxidation in acid solutions, as shown in the following equations:

$$S + 4Cu^{+2} + 2SO_4^{-2} + 3H_2O \rightarrow 4Cu^{+1} + 6H^{+1} + SO_3^{-2} + 2SO_4^{-2}, \qquad (3d)$$
$$\text{emf} = -0.45 - (-0.17) = -0.28 \text{ volt};$$

$$H_2SO_3 + H_2O + 2Cu^{+2} + 2SO_4^{-2} \rightarrow 2Cu^{+1} + 3SO_4^{-2} + 4H^{+1}, \qquad (3e)$$
$$\text{emf} = -0.20 - (-0.17) = -0.03 \text{ volt}.$$

Equations (3*d*) and (3*e*) show that, under standard conditions, the reduction of cupric ion would never achieve the first stage (reduction to cuprous ion). Because of this, the fact that the emf of the second stage of the copper reduction is less than any of the emf's in the sulfur oxidation has no practical importance because no cuprous ion would be formed to permit the cuprous-copper half-reaction to come into being.

Because the ferric ion reduction, however, proceeds as shown in equation (3*b*) with a net emf of +0.32 volt, the end-result is (when both ferric and cupric ions are present) a net positive emf in favor of both the second and the third stage of the sulfur oxidation under standard conditions. This emf will be only 0.04 volt in the second step in sulfur oxidation but will rise to 0.44 volt in the third. Even though conditions deviate from standard, as long as the concentrations of ferric and cupric ions are nearly equal or the latter are in the minority, the greater emf of the ferric ion reduction will overcome the weaker force of that of copper to cupric ion.

If it is remembered that the solutions which carry out both supergene and hypogene replacements almost certainly do not contain their various components in unit effective concentrations, it is apparent that the voltages given in the preceding paragraph are not immutable. With any change in the effective concentration of any of the components entering into the oxidation-reduction reaction in question, there will be a change in the reaction's net emf. Of the ions taking part in the reactions of the oxidation of sulfur, the one whose change most greatly varies the emf is the hydrogen. This can be seen from the Nernst equation given below. It is unlikely that supergene solutions, much less hypogene, will have an effective concentration of hydrogen ion greater than 10^{-3} (pH $= 3$). Accepting 10^{-3} as the effective concentration of the hydrogen ion in the half-reaction for the second stage of sulfide ion oxidation (the activities of the other components remaining at unity), a new oxidation potential for this half-reaction can be determined, as follows:

$$E = E^{\circ}_{25^{\circ}} - \frac{RT}{nF}\, ln\, \frac{a^{c}_{C}\, a^{d}_{D}}{a^{a}_{A}\, a^{b}_{B}} \qquad \text{(Nernst equation)}$$

$$= -0.45 - \frac{2.303RT}{nF} \log_{10} \frac{a_{SO_3} a^{4}_{H}}{a_{S}\, a^{3}_{H_2O}}$$

$$= -0.45 - \frac{0.05915}{4} \log_{10} \frac{1\,(10^{-3})^{4}}{1 \times 1}$$

$$= -0.45 - 0.0148 \log_{10} 10^{-12}$$

$$= -0.45 + 0.18 = -0.27 \text{ volt}\,.$$

Thus, in acid solutions of the maximum acidity to be expected under even supergene conditions, the oxidation potential of the half-reaction in question (and in any other in which the hydrogen ion is one of the reducible substances produced) is increased over that obtained with a hydrogen-ion activity of unity. In this particular case the oxidation potential is still less than that for the oxidation of cuprous to cupric ion. This means that the reaction of equation (3*d*) would not go forward unless again aided by ferric ion reduction. The amount of help required from ferric ion is, however, much less because the deficit in equation (3*d*) is, with the substitution of the new sulfur-sulfite potential, only 0.10 volt. This same change in hydrogen-ion concentration would also affect the sulfite-sulfate potential, changing it from −0.20 to −0.02 volt.

If the hydrogen-ion concentration were to be lowered even more, to 10^{-5}, for example, the oxidation potential for the sulfur-sulfite half-reaction would be raised to −0.15 volt and that of the sulfite-sulfate to +0.10 volt. Both, then, would be greater than those of either stage in the copper reactions because the change of hydrogen-ion concentration has no effect on the various copper potentials. In strongly acid solutions in which the reduction of cuprous ion to copper or cupric ion to cuprous or native copper is involved, it may be necessary to appeal to a stronger oxidizing agent than ferric iron to aid the particular copper ion reduction in question. Such an agent might be manganic (Mn^{+3}) ion, which, in its reduction to manganous (Mn^{+2}), has a reduction potential of 1.51 volts.

From this discussion it would seem reasonable that the reduction of cupric to cuprous ion would, in most weakly acid supergene solutions, insure the oxidation of sulfide ion through its three stages to the sexivalent positive sulfur of the sulfate radical. If the replacing solutions were to be alkaline, as they well may be under hypogene conditions, then the oxidation potentials for the sulfur half-reactions would be those for alkaline solutions and would have far higher values. This means that, in alkaline solutions, any half-reaction of an emf less than, say, +0.40 volt would have sufficient emf to drive S^{-2} through the three steps to S^{+6}. This widens the field of choice of oxidizing agents greatly over that which is available for use in supergene environments. It must be emphasized again that it is not claimed that replacing solutions have all ions but the hydrogen at unit effective concentrations, even though, for simplicity, they have been allowed to remain so in the calculations just made. But it is claimed that the Nernst equation shows that a change in hydrogen-ion concentration will have a far greater effect on the oxidation potential of the second and third stages of sulfide ion oxidation than will changes in the effective concentrations of the other components. This is true because the power to which the hydrogen-ion concentration is raised in that equation is higher than for any other constituent. This idea can be made more general in application by saying that a change in the effective concentration of the hydrogen ion will have a greater effect than will a change in concentration of any other ion on the value of the oxidation potential of any half-reaction in which the number of hydrogen ions formed by it is greater than the number of ions of any other reducible substance produced.

HOST IN EXCESS

In the replacement reaction (sphalerite by argentite) which has just been discussed, it was necessary to remove and oxidize some sulfide ions from the host because an equal volume of the guest could not utilize them. The reverse of this situation would obtain if the volume of the host being replaced held less sulfur than would be required by an equal volume of guest. This type of metasomatism is illustrated by the replacement of sphalerite by covellite.

$$ZnS + Cu^{+2} + SO_4^{-2} \rightarrow CuS + Zn^{+2} + SO_4^{-2}\,. \qquad (4a)$$

The actual space occupied by 1 formula unit (molecule) of sphalerite is 39.37 $Å^3$ (see table 1), while that taken up by 1 molecule of covellite is 34.24 $Å^3$. If replacement of sphalerite by covellite occurred exactly as represented by equation (4*a*), there would be a loss of nearly 15 per cent in volume during the process. Because such a loss in volume does not take place under the conditions of hypogene or supergene metasomatism, it follows that equation (4*a*) must be modified to equate not only molecular units but unit volumes as well. From the data just given it is quickly seen that the replacement of 1 molecule of sphalerite (Vol. = 39.37 $Å^3$) by 1 molecule of covellite (Vol. = 34.24 $Å^3$) would leave a deficit of 5.13 $Å^3$ of covellite if volume-for-volume relations are to be preserved. These 5.13 $Å^3$ are almost exactly 15 per cent of the volume of 1 formula unit of covellite. This means that 1.15 formula units of covellite must be made available for each formula unit of sphalerite replaced if no loss in volume is to occur. To eliminate fractions of molecules, it is necessary to multiply the 1.15 molecules of covellite and the 1 molecule of sphalerite by 20, i.e., 23 molecules of covellite must be provided for each 20 molecules of sphalerite removed. Where, however, are these extra 3 molecules of covellite to be obtained? They cannot be secured by the replacement of zinc atoms from additional sphalerite by cupric ions from the replacing solution because that would further widen the gap between initial and final volumes. There is, it must be emphasized, a lack not merely of cupric ions but of sulfide ions as well, and these sulfide ions must come from the solutions involved and not from the body of the host mineral. In short, it is not sufficient to introduce three additional cupric ions from a solution in which they are in balance with some acid radical or ion other than the sulfide. Each of the 3 cupric ions must be, at the time of entry into the host volume, in balance with a sulfide ion, and all 3 molecules must be introduced into the space vacated by the 20 zinc atoms at the same time as are the 20 cupric ions required to balance the 20 sulfide ions inherited from the sphalerite.

The reaction which has just been discussed probably can be seen more clearly if it is expressed in the form of an equation, as it is in equation (5*a*), which follows:

$$\left.\begin{array}{l} \underset{(787.40\ \text{Å}^3)}{20\text{ZnS}} + 23\text{Cu}^{+2} + 20\text{SO}_4^{-2} + 3\text{S}^{-2} \rightarrow \underset{(787.52\ \text{Å}^3)}{23\text{CuS}} + 20\text{Zn}^{+2} + 20\text{SO}_4^{-2}\,, \end{array}\right\} \qquad (5a)$$

Volume difference = 0.02 per cent .

Equation (5*a*), however, indicates that the additional sulfur needed to put the equation on a volume-for-volume basis is transported in the replacing solutions as sulfide ion. If this sulfide ion is presumed to come from hydrogen sulfide, the work of Allen, Crenshaw, and Johnson (1912, p. 172) shows that any solutions containing it must have been essentially lacking in oxygen. If such an environment did not obtain, the hydrogen sulfide would be rapidly oxidized as follows:

$$H_2S + 2O_2 = H_2SO_4\,.$$

This hydrogen sulfide also must be in excess over ferric ion, or the iron would be reduced to the ferrous state and all the sulfide ion oxidized in steps (see eqs. [3*a*], [3*b*], and [3*c*]) to the S^{+6} of the sulfate radical. It is doubtful if these requirements could be fulfilled on any appreciable scale above the water table. Below it, however, it seems probable that both conditions could be met in considerable volumes of supergene enriching solutions (Emmons, 1917, pp. 98–99 and 150–152). Granting this, equation (5*a*) can be modified slightly to show the presence of H_2S without affecting the volumetric balance it attained:

$$20\text{ZnS} + 23\text{Cu}^{+2} + 23\text{SO}_4^{-2} + 3\text{H}_2\text{S} \rightarrow 23\text{CuS} + 20\text{Zn}^{+2} + 20\text{SO}_4^{-2} + 6\text{H}^{+1} + 3\text{SO}_4^{-2}\,. \qquad (5b)$$

The additional covellite required for volumetric balance, then, is provided by a direct reaction between sulfide ion from the dissociation of hydrogen sulfide and cupric ion, as is shown in equation (4*b*) as follows:

$$\text{Cu}^{+2} + \text{SO}_4^{-2} + \text{H}_2\text{S} \rightarrow \text{CuS} + 2\text{H}^{+1} + \text{SO}_4^{-2}\,. \qquad (4b)$$

Equation (5*b*) is equation (4*a*) multiplied by 20 and added to equation (4*b*) multiplied by 3.

Because no oxidation or reduction is involved in equation (5*b*), it is necessary to consider the free-energy relations concerned, to see whether the reaction will go forward. This is easily done by examining the free energies of formation of the products and reactants in the two component equations of equation (5*b*). From Kelley's (1937) and Verhoogen's (1938*a*) calculations of the free energies of formation of metal ions, sulfides, and sulfates (see table 3), it can be determined that both these reactions will go to the right when carried out separately. The free-energy change in (4*a*) in calories is:

TABLE 3*

Compound or Ion	Free Energy of Formation in Calories at 298.1° K.	Compound or Ion	Free Energy of Formation in Calories at 298.1° K.
$MnSO_4$	−228,020	CdS	− 33,490
$ZnSO_4$	−208,090	FeS_a	− 23,390
$CdSO_4$	−195,760	PbS	− 22,730
$PbSO_4$	−193,650	CoS	− 21,060
$CoSO_4$	−183,650	Cu_2S	− 20,550
H_2SO_4	−176,540	Fe^{+2}	− 20,240
SO_4^{-2}	−176,540	Co^{+2}	− 13,375
$CuSO_4$	−157,950	CuS	− 11,700
Ag_2SO_4	−146,800	Ag_2S	− 8,680
MoS_2	− 54,190	$H_2S_{(g)}$	− 7,865
Mn^{+2}	− 50,730	$H_2S_{(aq)}$	− 6,490
MnS	− 45,520	Pb^{+2}	− 5,810
FeS_2	− 44,720	HgS	− 3,350 (+above 200°)
ZnS	− 40,370	Cu^{+2}	+ 15,900
Bi_2S_3	− 39,140		
Zn^{+2}	− 35,100		

* The data from which this table was compiled were derived from Kelley (1937) in the main, but values were calculated for a few metal ions not given by him, using the formula given by Verhoogen (1938*a*) as his equation (12).

REACTANTS		PRODUCTS	
ZnS	−40,370	CuS	−11,700
Cu^{+2}	+15,900	Zn^{+2}	−35,100
	−24,470		−46,800
			−24,470
			−22,330

In equation (4*b*) the change in calories is:

REACTANTS		PRODUCTS	
Cu^{+2}	+15,900	CuS	−11,700
$H_2S_{(aq)}$	− 6,490		+ 9,410
	+ 9,410		−21,110

Thus not only is the free-energy change such that the reactions will tend to go as written, but also it is nearly the same in each of the two cases. It seems, therefore, quite likely that both reactions could be carried out simultaneously in supergene solutions, as is suggested in equation (5*b*).

The source of the necessary hydrogen sulfide used in reactions such as equation (5*b*) probably is the reaction of sulfuric acid (or rather the hydrogen ion produced by the ionization of sulfuric acid) with such sulfides as pyrrhotite, sphalerite, alabandite, and galena. From a consideration of the free energies involved, it would be doubtful whether galena would contribute any hydrogen sulfide, except that the products probably include $PbSO_{4(s)}$ instead of Pb^{+2} and SO_4^{-2}. Wells, quoted by Emmons (1917, pp. 119–121), found that some hydrogen sulfide was produced when galena was treated with dilute sulfuric acid, but only about one-fourth as much as was formed when sphalerite was similarly attacked. This replacement of lead by hydrogen is consistent with the position of lead in the electromotive series and with free-energy change, as can be seen in the following equation:

$$PbS + 2H^{+1} + SO_4^{-2} \rightarrow H_2S + PbSO_4 .$$

REACTANTS		PRODUCTS	
PbS	− 22,730 cal.	$PbSO_4$	−193,650 cal.
H_2SO_4	−176,540	H_2S	− 6,490
	−199,270		−200,140
			−199,270
			− 870

As can be seen from table 3, two of the other three sulfides (MnS and ZnS) have free energies of formation low enough to permit the formation of H_2S when they are brought into contact with dilute sulfuric acid. Pyrrhotite is of particular interest because it is known that it produces far more hydrogen sulfide in this manner than any other sulfide does. A quantitative examination of that reaction, however, cannot be made, since nothing is known about its free energy of formation. If it is assumed that its free energy is about that of ferrous sulfide, then the large production of H_2S can be explained. The reaction is as follows:

$$FeS_a + 2H^{+1} + SO_4^{-2} \rightarrow H_2S + Fe^{+2} + SO_4^{-2} .$$

REACTANTS		PRODUCTS	
FeS_a	−23,390 cal.	Fe^{+2}	−20,240 cal.
		H_2S	− 6,490
			−26,730
			−23,390
			− 3,340

Pyrite is a special case, since its very low free energy of formation, much lower than would be expected, prevents it from reacting with sulfuric acid as does pyrrhotite.

It must be understood that the values used for the free energies of formation are exact only for 1 atm. of pressure and 298.1° K. Under any other conditions the values here used are not correct but probably indicate the general order of magnitude, in any environment, of the free energies of formation of the various compounds and ions considered. A comprehensive study of the problem is made by Verhoogen (1938*a* and *b*).

As an alternate scheme to that just proposed, it is suggested that there is a possibility that the sulfur is brought to the scene of the metasomatism as the sexivalent positive sulfur of the sulfate radical and there reduced to S^{-2} by the action of some naturally occurring reducing agent (of higher oxidation potential than that of any of the three half-reactions pertaining to the reduction of S^{+6}). From table 2, it can be seen that such a reducing agent must have an oxidation potential somewhat greater than −0.14 volt, preferably 0.00 volt or higher. A reducing agent which meets these requirements and which also is likely to be found in supergene enriching solutions is not easily discovered in the usual oxidation-reduction tables. The best suggestion that I have been able to find is the oxidation of one or more of the following ions: Cr^{+2} to Cr^{+3}, V^{+2} to V^{+3}, Mo^{+3} to Mo^{+5}, Ti^{+2} to Ti^{+3}, Cd^{+1} to Cd^{+2}, or any similar reaction for which a standard potential can be discovered. Since cadmium is to be found associated with most sphalerite, it is being used to write a possible equation, even though it is realized that the rather uncommon Cd^{+1} ion probably seldom is available in sufficient quantity to do the job. The possible equation is:

$$\left.\begin{aligned}&20ZnS + 23Cu^{+2} + 23SO_4^{-2} + 24Cd^{+1} + 12SO_4^{-2} + 24H^{+1} + 12SO_4^{-2} \rightarrow \\ &\qquad 23CuS + 20Zn^{+2} + 20SO_4^{-2} + 24Cd^{+2} + 24SO_4^{-2} + 12H_2O\,.\end{aligned}\right\} (5c)$$

If the rather unusual Cd_2^{+2} ion is not thought likely to be present or, if present, to be in sufficient abundance to accomplish the S^{+6} reduction, it may be substituted for, or supplemented by, ions of one or more metals of similar chemical properties and oxidation potentials, as follows:

$$\left.\begin{aligned}&20ZnS + 23Cu^{+2} + 23SO_4^{-2} + 24Cr^{+2} + 24SO_4^{-2} + 24H^{+1} + 12SO_4^{-2} \rightarrow \\ &\qquad 23CuS + 20Zn^{+2} + 20SO_4^{-2} + 24Cr^{+3} + 36SO_4^{-2} + 12H_2O\,.\end{aligned}\right\} (5d)$$

In addition to the suitable metals, potentials for the oxidation of whose ions are given in table 2, others which should be satisfactory include tungsten, zirconium, and tantalum. The potentials for the half-reactions of the ions of these metals are not available but should be within the range in potentials from the Cd_2^{+2} oxidation to that of Mo^{+3}(r).

It is chemically axiomatic that the ferric-sulfide ion reduction oxidation of equations (2*a*) and (2*b*), on the one hand, and the hydrogen sulfide precipitation of equation (5*b*) or the sulfate ion reduction used in equations (5*c*) and (5*d*), on the other, cannot go on simultaneously in the same system. A sulfide ion cannot be oxidized under the same conditions as those under which a sulfate ion is being reduced. Nor can the precipitation of sulfides by H_2S go on in the presence of ferric ion. In the open systems which exist in ground waters in contact with ore bodies, however, it is possible that, at different times, both these types of mutually exclusive reactions may take place in the same area or that, at the same time, both may be going forward in two different places not necessarily removed from each other by any great distance.

If the reasoning of the last paragraph is correct, it follows that reactions of the type of equations (2*a*) and (2*b*) cannot occur in supergene processes at the same time and in the same place as those like equations (5*b*), (5*c*), and (5*d*). It also seems apparent that in the rare simultaneous supergene replacements of one mineral by two or more, volume adjustments must be made either through the deposition of the replacing minerals in proper proportions to achieve volumetric balance or through the help of one of the two general types of supplementary reactions, but not through the aid of both. For example, a molecularly balanced equation can be written for the replacement of sphalerite by covellite and argentite, unlikely as that would be under supergene conditions, using any desired ratio between the replacing minerals. But if such an equation is to be balanced in relation to volume as well, the ratio of covellite to argentite must be 3 to 1, and even then the volume difference is over 2 per cent. A volume difference of about 0.5 per cent can be attained (see eq. [6]) if 7 molecules of sphalerite are used, 3 being replaced by covellite, 3 by argentite, and 1 being dissolved by the action of ferric ion to oxidize the sulfur of one sphalerite molecule to sulfate ion, as follows:

$$\left.\begin{aligned}&\underset{(275.59\ \text{Å}^3)}{7ZnS} + 6Ag^{+1} + 3Cu^{+2} + 8Fe^{+3} + 18SO_4^{-2} + 4H_2O \rightarrow\\ &\qquad \underset{(174.33\ \text{Å}^3)}{3Ag_2S} + \underset{(102.72\ \text{Å}^3)}{3CuS} + 7Zn^{+2} + 8Fe^{+2} + 8H^{+1} + 19SO_4^{-2}\,,\end{aligned}\right\}(6)$$

$$\text{Volume difference} = 0.53 \text{ per cent}\,.$$

It seems probable that there are definite limits as to the ratios in which any two or more minerals can replace a host, under hypogene as well as supergene conditions. These limits are set, it appears, by the relative volumes of the formula units of the minerals concerned. In some cases volumetric balance may be attained without recourse to the dissolving of additional host or the precipitation of extra amounts of one or more of the guests. It seems likely, however, that one or the other of the two processes (additional solution or deposition) must be invoked in most cases of such reactions. Both cannot take place at the same time and in the same system.

FURTHER EXAMPLES

Further to illustrate the volumetric and molecular balancing of replacement equations, the following eight reactions have been selected from those which are generally accepted as occurring during supergene metasomatism.[5] From the data in table 1 it appears that some of these reactions come near enough to volumetric balance not to need any modification, but the others require a good deal.

$$CuFeS_2 + Cu^{+2} + SO_4^{-2} \rightarrow 2CuS + Fe^{+2} + SO_4^{-2}\,, \tag{7}$$

$$5FeS_2 + 14Cu^{+2} + 14SO_4^{-2} + 12H_2O \rightarrow 7Cu_2S + 5Fe^{+2} + 5SO_4^{-2} + 24H^{+1} + 12SO_4^{-2}\,, \tag{8a}$$

$$4FeS_2 + 7Cu^{+2} + 7SO_4^{-2} + 4H_2O \rightarrow 7CuS + 4Fe^{+2} + 4SO_4^{-2} + 8H^{+1} + 4SO_4^{-2}\,, \tag{9a}$$

$$Cu_5FeS_4 + Cu^{+2} + SO_4^{-2} \rightarrow 2Cu_2S + 2CuS + Fe^{+2} + SO_4^{-2}\,, \tag{10}$$

$$5CuS + 3Cu^{+2} + 3SO_4^{-2} + 4H_2O \rightarrow 4Cu_2S + 8H^{+1} + 4SO_4^{-2}\,, \tag{11a}$$

$$Cu_2S + 4Ag^{+1} + 2SO_4^{-2} \rightarrow 2Cu^{+2} + 2SO_4^{-2} + Ag_2S + 2Ag\,, \tag{12a}$$

$$Cu_2S + 6Fe^{+3} + 9SO_4^{-2} + 4H_2O \rightarrow 2Cu + 6Fe^{+2} + 6SO_4^{-2} + 8H^{+1} + 4SO_4^{-2}\,, \tag{13a}$$

$$PbS + 4O \rightarrow PbSO_4\,. \tag{14a}$$

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[5] Eqs. (7)–(11a) have been taken from Bateman (1942, p. 277), (12) from Lindgren (1905, p. 101), (13a) from Emmons (1917, p. 258), and (14a) from Anderson (1930, p. 529).

Inspection of the equations just given and the unit volumes of the host and guest minerals show that equations (7) and (10) come fairly close to volumetric balance without any change in proportions or reactants. Equation (7) balances within 3.1 per cent and equation (10) within 2.2 per cent; although these deviations are somewhat higher than the limit of 1 to 2 parts in 100 for such variations, they are allowed to stand without any rearrangement.

Equations (8*a*), (9*a*), (11*a*), (12*a*), (13*a*), and (14*a*), however, require considerable change to put them in both volumetric and molecular balance. In equations (8*a*) and (9*a*) it is necessary to remove more pyrite than the reaction given will permit. This means that another reaction which dissolves pyrite and leaves no solid products behind must go on concurrently with those of equations (8*a*) and (9*a*). The reactions suggested by Stokes (1901, p. 15) and given as equations (15*a*) and (16) probably will do the job. Equation (15*a*) is quite similar to equation (3*a*), and equation (16) is the total of equations (3*b*) and (3*c*) multiplied by 2.

$$FeS_2 + 2Fe^{+3} + 3SO_4^{-2} \rightarrow 3Fe^{+2} + 3SO_4^{-2} + 2S\,, \tag{15a}$$

$$2S + 12Fe^{+3} + 18SO_4^{-2} + 8H_2O \rightarrow 12Fe^{+2} + 12SO_4^{-2} + 16H^{+1} + 8SO_4^{-2}\,. \tag{16}$$

If these two reactions are combined with equations (8*a*) and (9*a*), respectively, with the necessary pyrite added to achieve volumetric balance, the following equations, (8*b*) and (9*b*), result:

$$\left.\begin{aligned}&\underset{(315.84\ \text{Å}^3)}{8FeS_2} + 42Fe^{+3} + 63SO_4^{-2} + 14Cu^{+2} + 14SO_4^{-2} + 36H_2O \rightarrow\\ &\qquad \underset{(318.78\ \text{Å}^3)}{7Cu_2S} + 50Fe^{+2} + 50SO_4^{-2} + 72H^{+1} + 36SO_4^{-2}\,,\end{aligned}\right\}(8b)$$

Volume difference = 0.92 per cent ;

$$\left.\begin{aligned}&\underset{(236.88\ \text{Å}^3)}{6FeS_2} + 28Fe^{+3} + 42SO_4^{-2} + 7Cu^{+2} + 7SO_4^{-2} + 20H_2O \rightarrow\\ &\qquad \underset{(239.68\ \text{Å}^3)}{7CuS} + 34Fe^{+2} + 34SO_4^{-2} + 40H^{+1} + 20SO_4^{-2}\,,\end{aligned}\right\}(9b)$$

Volume difference = 1.16 per cent .

The modification of equation (11*a*) presents a somewhat different problem because here it is necessary to remove more covellite than the reaction given will permit, to provide room for the chalcocite formed. To accomplish this, a second reaction must go on at the same time as that of equation (11*a*) to remove the excess covellite without leaving any solid products behind. A slight change in equation (15*a*) should take care of this, however, while equation (16) must be halved. This change in equation (15*a*) is given in equation (15*b*):

$$CuS + 2Fe^{+3} + 3SO_4^{-2} \rightarrow Cu^{+2} + SO_4^{-2} + 2Fe^{+2} + 2SO_4^{-2} + S\,. \tag{15b}$$

When equations (11*a*), (15*b*), and (16) are combined, the result is:

$$\left.\begin{aligned}&\underset{(719.04\ \text{Å}^3)}{21\text{CuS}} + 12\text{Cu}^{+2} + 12\text{SO}_4^{-2} + 8\text{Fe}^{+3} + 12\text{SO}_4^{-2} + 20\text{H}_2\text{O} \rightarrow \\ &\qquad\qquad \underset{(728.64\ \text{Å}^3)}{16\text{Cu}_2\text{S}} + \text{Cu}^{+2} + \text{SO}_4^{-2} + 8\text{Fe}^{+2} + 8\text{SO}_4^{-2} + 40\text{H}^{+1} + 20\text{SO}_4^{-2}\,,\end{aligned}\right\} \quad (11b)$$

Volume difference = 1.32 per cent.

To adapt equation (12*a*) requires a further change in equation (15*a*) and its combination with equation (16), which again it is necessary to halve:

$$\text{Cu}_2\text{S} + 4\text{Fe}^{+3} + 6\text{SO}_4^{-2} \rightarrow 2\text{Cu}^{+2} + 2\text{SO}_4^{-2} + 4\text{Fe}^{+2} + 4\text{SO}_4^{-2} + \text{S}\,. \qquad (15c)$$

In equation (15*c*), in addition to oxidizing the sulfide ion, ferric ion also is called upon to oxidize cuprous ion as well. From table 2 it can be seen that the necessary electronic reactions are such that the reaction will go with an emf of +0.60 volt:

$$\left.\begin{aligned}&\underset{(91.08\ \text{Å}^3)}{2\text{Cu}_2\text{S}} + 4\text{Ag}^{+1} + 2\text{SO}_4^{-2} + 10\text{Fe}^{+3} + 15\text{SO}_4^{-2} + 4\text{H}_2\text{O} \rightarrow \\ &\qquad \underset{(58.11\ \text{Å}^3)}{\text{Ag}_2\text{S}} + \underset{(33.90\ \text{Å}^3)}{2\text{Ag}} + 4\text{Cu}^{+2} + 4\text{SO}_4^{-2} + 10\text{Fe}^{+2} + 10\text{SO}_4^{-2} + 8\text{H}^{+1} + 4\text{SO}_4^{-2}\,,\end{aligned}\right\} \quad (12b)$$

Volume difference = 1.01 per cent.

In equation (12*b*) ferric ion is assisted in its oxidation of the sulfide ion by the reduction of both cuprous and silver ions to the native metals. From table 2 it is apparent that the emf of the half-reaction involving silver is even less than that involving ferrous ion. Therefore, silver ion can aid in all three stages of sulfide ion oxidation to an even greater extent than can ferric. The emf of the half-reaction containing the cuprous ion, although greater than that with ferrous ion, still is less than that of the second stage in the oxidation of sulfide ion and, therefore, will aid in all stages of the oxidation.

The rearrangement of equation (13*a*) is somewhat more complicated than is that of those from equations (7) through (12*a*). It resembles equation (4*a*) in that, if molecularly balanced, the volume of solid reactant is larger than that of solid product. It differs from equation (4*a*), however, in that the reduction needed for volumetric balance is not that of sulfate to sulfide ion but that of cupric ion to native copper. The reduction of cuprous ion from the chalcocite to copper must also be taken care of, as well as the oxidation of the sulfide ion of the chalcocite to sulfate. It has already been pointed out in the discussion of equation (12*b*) that the reduction of cuprous ion to native copper can aid in all stages of the oxidation of sulfide ion. This is not true of the reduction of cupric ion to native copper, however, because that half-reaction has an emf of −0.345 volt, which is greater than that for the second stage of sulfide oxidation. Under standard conditions, therefore, the reduction of cupric ion to native copper would aid only in the first and third stages of the oxidation and would, presumably, oppose the reducing action of ferric ion in the second. If the cupric ion reduction took place in stages, from cupric to cuprous and cuprous to native copper, as it probably does, the relatively high oxidation potential of the first stage of copper reduction would prevent its aiding in both the second and the third stages of sulfide oxidation.

The second stage of reduction (cuprous to copper) would create an emf sufficient to take part in all stages of the oxidation. The initial modification of equation (13*a*) on this basis is as follows:

$$\left.\begin{aligned}&Cu_2S + 2Cu^{+2} + 2SO_4^{-2} + 2Fe^{+3} + 3SO_4^{-2} + 4H_2O \rightarrow \\ &\qquad 4Cu + 2Fe^{+2} + 2SO_4^{-2} + 8H^{+1} + 4SO_4^{-2}\,.\end{aligned}\right\} \quad (13b)$$

Equation (13*b*), although in molecular and electrical balance, still is unbalanced volumetrically by about 3 per cent. This unbalance can be largely removed by combining equation (13*b*), multiplied by 12, with equation (13*a*), taken only once. This combination gives an exact molecular and electrical balance and is within less than 1 per cent of being volumetrically balanced, as can be seen from equation (13*c*):

$$\left.\begin{aligned}&\underset{(592.02\ \text{Å}^3)}{13Cu_2S} + 24Cu^{+2} + 24SO_4^{-2} + 30Fe^{+3} + 45SO_4^{-2} + 52H_2O \rightarrow \\ &\qquad \underset{(587.00\ \text{Å}^3)}{50Cu} + 30Fe^{+2} + 30SO_4^{-2} + 104H^{+1} + 52SO_4^{-2}\,,\end{aligned}\right\} \quad (13c)$$

Volume difference = 0.85 per cent.

The adaptation of equation (14*a*) to the requirements of volumetric balance is relatively simple, but the explanation of why one-third of the lead sulfate formed is removed in solution is not:

$$\left.\underset{(156.39\ \text{Å}^3)}{3PbS} + 12O \rightarrow \underset{(156.96\ \text{Å}^3)}{2PbSO_4} + Pb^{+2} + SO_4^{-2}\,,\right\} \quad (14b)$$

Volume difference = 0.36 per cent.

SULFIDE AND SULFOSALT REPLACEMENTS—HYPOGENE CONDITIONS

GENERAL

The writing of equations for the reactions of hypogene replacements is a somewhat different and more difficult task than for those of the supergene zone. It is possible, in the latter type of reactions, to put in the equation what well may be all the essential constituents of the reaction, both those in the solid state and those in solution. Enough is known of the chemistry of supergene replacements to do this without great possibility of error. Such completeness is not, however, to be attained in the equational expression of hypogene metasomatism. There is too much uncertainty as to the manner in which the essential and additional ions needed for the reaction are brought to the scene. There is even less knowledge of those ions which are in chemical balance with those which are definitely present in the replacing solutions but which leave no trace of their presence in the replacement minerals. It is beyond the scope of this paper, for example, to debate the merits of colloidal versus complex ion transfer. It is apparent, as well, that complete equations for hypogene replacements cannot be written until that controversy is settled and more is known of the actual composition of, as well as manner of transportation in, the ore-forming fluid. Nevertheless, the possibility remains that skeleton equations, expressing the basic changes which take place, can be written for hypogene replacement reactions.

In completely balanced supergene equations it is possible to determine whether extra replacing mineral must be formed from the components of the ore solutions or additional amounts of the replaced mineral dissolved over the quantities required for simple molecular balance. This is not true of their hypogene counterparts, however, because simple molecularly balanced equations are not possible with the limited data at hand. All that it is possible to do is, by division of one formula unit volume by the other (or others), to determine the number of molecules of guest needed to equal volumetrically a definite number of host molecules. If the correct molecular quantities of the two or more minerals are set forth on opposite sides of an equation, it is immediately apparent that certain components must be added on both sides of the equation. Those on the left are the additional ions needed, over those furnished by the host, to create the guest. Those on the right are those present in the host but not needed in the formation of the guest.

The problem of addition and subtraction of metal ions only is negligible. But, if sulfur is involved, its addition or subtraction is less easily explained. If more sulfide ion is furnished by the host than can be used in the guest, it seems probable that the excess sulfide ion must be oxidized to sulfate for ready removal from the site of the metasomatism. If, on the other hand, less sulfur ion is provided by the host than is needed by the guest, sulfide ion must be supplied from the ore-forming fluid. The mechanism for the production of the requisite amount of sulfide ion is highly debatable. It may come from sulfide ion carried in solution in balance with a free cation, from the dissociation of a metal-sulfur complex ion, or from a colloidal dispersion of metal sulfides. So far as this study is concerned, this supplementary sulfide ion will be assumed to be available when needed as sulfide ion, and no attempt will be made to say how or in what state it reached the point at which it combines with the ions of the host to form the guest. Finally, there are rare cases, such as are indicated by equations (17) and (23), in which sulfide ion is present in the host in the exact quantities needed by the guest. Such happy coincidences are, unfortunately, rare.

Thus in hypogene, as in supergene, replacements there are three general types of molecularly and volumetrically balanced equations: (1) those in which more sulfide ion is provided by the guest than the host can use, which necessitates oxidation and removal in solution of the unwanted sulfide; (2) those in which less sulfide ion is furnished by the host than is requisite to the formation of the guest, which requires the addition of the needed sulfide ion from the solutions; and (3) those in which sulfide ion is present in the host in exactly the amounts which are called for by the guest.

EXAMPLES

The different methods of approach to the balancing of these three types of equations are shown in equations (17)–(26), inclusive, which follow. They were selected to show as wide a variety of hypogene replacements as can be managed from such a small number of examples. They show what can be done to balance such reactions as to both weight and volume.

$$\left.\begin{array}{l}\underset{(573.43\ \text{Å}^3)}{11\mathrm{PbS}} + 8\mathrm{Ag}^{+1} + 8\mathrm{Cu}^{+1} + 2\mathrm{Sb}^{+3} \longrightarrow \underset{(582.11\ \text{Å}^3)}{\mathrm{Ag_8Cu_8Sb_2S_{11}}} + 11\mathrm{Pb}^{+2}\,,\end{array}\right\} \qquad (17)$$

Volume difference = 1.47 per cent.

It is quite possible that the copper in the solution accomplishing the replacement of galena by polybasite is in the cupric state and is reduced to the cuprous during precipitation, but there are no data on which to base such a conclusion except the relative instability of the cuprous ion vis-à-vis the cupric:

$$\left.\begin{array}{l}\underset{(591.00\ \text{Å}^3)}{15ZnS} + 4Pb^{+2} + 8Fe^{+3} + 6Sb^{+3} + 4H_2O \rightarrow \underset{(582.11\ \text{Å}^3)}{Pb_4FeSb_6S_{14}} + 15Zn^{+2} + SO_4^{-2} \\ \qquad + 7Fe^{+2} + 8H^{+1},\end{array}\right\}(18)$$

Volume difference = 1.11 per cent.

In this replacement of sphalerite by jamesonite, the presence of an additional molecule of sphalerite, above that required for molecular balance, necessitates the presence in the solution of iron in the ferric state to convert the sulfide ion to sulfate:

$$\left.\underset{(565.60\ \text{Å}^3)}{8CuFeS_2} + 6Cu^{+2} + 7Sn^{+2} + 12S^{-2} \rightarrow \underset{(559.51\ \text{Å}^3)}{7Cu_2FeSnS_4} + Fe^{+2},\right\}(19)$$

Volume difference = 1.08 per cent.

It will be noted in this metasomatism of chalcopyrite by stannite that the conversion of the cupric copper to cuprous is assumed to be accomplished by the oxidation of the stannous ion from the solution. From table 2 it can be seen that this reaction will go in the direction indicated but slowly because the net emf will be but +0.02 volt under standard conditions.

$$\left.\begin{array}{l}\underset{(551.19\ \text{Å}^3)}{Cu_{12}Sb_4S_{13}} + 14Fe^{+3} + 4Pb^{+2} + 4H_2O \rightarrow \underset{(543.00\ \text{Å}^3)}{4PbCuSbS_3} + 8Cu^{+2} \\ \qquad + 14Fe^{+2} + SO_4^{-2} + 8H^{+1},\end{array}\right\}(20)$$

Volume difference = 1.47 per cent.

This replacement of tetrahedrite by bournonite requires the oxidation of one ion of sulfur to the sexivalent sulfur of the sulfate ion, and the reduction of ferric to ferrous ion is used as the mechanism to achieve this. There is no reason why this reaction should not go forward as easily under hypogene as under supergene conditions, and, as has been pointed out on page 528, there are several reasons why it may proceed with a greater net emf:

$$\left.\begin{array}{l}\underset{(593.26\ \text{Å}^3)}{4Cu_3AsS_4} + Cu^{+2} + 2Fe^{+3} + 4H_2O \rightarrow \underset{(529.03\ \text{Å}^3)}{Cu_{12}As_4S_{13}} + \underset{(70.70\ \text{Å}^3)}{CuFeS_2} + Fe^{+2} \\ \qquad + SO_4^{-2} + 8H^{+1},\end{array}\right\}(21)$$

Volume difference = 1.06 per cent.

Equation (21) expresses the essential facts of the replacement of enargite by tennantite and chalcopyrite if the proportions of the latter two minerals to each other are correct, as their volume relations would seem to indicate. As written, the equation is balanced whether the arsenic of the enargite is considered to be tervalent or quinquevalent. The oxidation-reduction reactions taking place would be quite different, depending on the valence used, and both possibilities must be considered.

If the arsenic is As^{+5}, then not only must 1 sulfide ion (which the host contains in excess of the requirements of the guests) be oxidized to sexivalent positive sulfur, but 4 arsenic ions must be reduced to As^{+3}. In addition, 2 cuprous ions from the enargite must be oxidized to cupric copper. The latter oxidation can be accomplished by the simultaneous reduction of 2 ferric ions to the ferrous state and need concern us no further. The oxidation reduction of S^{-2} and As^{+5}, respectively, might be mutually accomplished if the emf's of the half-reactions taking place bear the proper voltage relations to one another, since the total loss and gain of electrons is equal. Unfortunately, I have been unable to find any published oxidation potentials for half-reactions which could, conceivably, be involved, such as,

$$As^{+3} \rightarrow As^{+5} + 2e^{-1}$$

or

$$2S^{-2} + As_2S_3 \rightarrow As_2S_5 + 2e^{-1} .$$

Nor have I been able to find the necessary data to permit approximate calculations of such potentials. For the reaction to go in the direction indicated in equation (21), however, the value of the arsenic half-reaction must be, under standard conditions, less than +0.48 volt, which is the lowest voltage produced by any of the half-reactions in the oxidation of S^{-2} to S^{+6} of the sulfate radical in the alkaline solutions which probably obtain under hypogene conditions.

If, on the other hand, the arsenic is As^{+3} in the enargite, the oxidation-reduction relations are quite different and must be considered in detail. One sulfide ion must be oxidized to sexivalent positive sulfur. As the reaction here is written, 6 of the 8 electrons removed from that sulfide ion are taken up by the reduction of an equal number of cupric to cuprous ions. The 2 remaining electrons are added to the 2 ferric ions, converting them to ferrous. If this reaction were attempted under the standard conditions at which the oxidation potentials in table 2 were determined, it almost certainly would not proceed as shown. The emf of the cuprous-cupric half-reaction is a −0.17 volt, while that of the second stage of the sulfide ion oxidation is −0.45 volt. This means that the reaction would tend to go in the opposite direction, with a net emf of +0.28 volt. This result might be avoided in neutral or alkaline solutions, as discussed on pages 528–529, in which the oxidation potential of the sulfur-sulfurous ion half-reaction would be greatly increased. Another possible solution to the problem would be to use another oxidizing agent of a considerably lower oxidation potential than that of the ferrous-ferric ion half-reaction. Such an oxidizing agent could be the reduction of manganic to manganous ion.

Equation (21), for the replacement of enargite by tennantite and chalcopyrite, probably represents only an intermediate stage in a reaction whose final product is chalcopyrite alone. All the original enargite and the subsequently developed tennantite are finally replaced by chalcopyrite. Perry (1933, p. 17) mentions the alteration of enargite to chalcopyrite with the formation of some tennantite, apparently as an intermediate step in the reaction, in the northeastern part of the Butte district. Ultimately, this tennantite, too, probably would have been converted to chalcopyrite, had the flow of solutions continued longer. The equation for the development of chalcopyrite from tennantite is given in equation (22):

$$\left.\begin{array}{l} 2Cu_{12}As_4S_{13} + 20Fe^{+3} + 4S^{-2} \rightarrow 15CuFeS_2 + 9Cu^{+2} + 8As^{+3} + 5Fe^{+2} , \\ (1058.08\ \text{Å}^3) \qquad\qquad\qquad\qquad (1060.50\ \text{Å}^3) \end{array}\right\} \quad (22)$$

Volume difference = 0.23 per cent .

In this reaction the agent used to oxidize the cuprous copper of the tennantite to the cupric of chalcopyrite is considered to be the ferric to ferrous iron reduction. Table 2 shows that this reaction will be, under standard conditions, driven forward with a net emf of +0.60 volt. As has already been stated, no attempt is made to account for the presence of sulfide ion in the replacing solutions in this reaction or that of equation (26), beyond the general one given on page 538.

The equation for the replacement of bornite by chalcopyrite and chalcocite, which follows, is one of the rare replacements, if it can properly be called such, in which no material is added from solution and none taken away. It well may be that such molecular rearrangements which require no extra ions to be given or taken away by the ore-forming fluid should not be called "replacements" at all, even though they are not recognized as unmixings from solid solution.

$$\left.\begin{array}{ccccc} Cu_5FeS_4 & \rightarrow & CuFeS_2 & + & 2Cu_2S\,, \\ (163.21\ \text{Å}^3) & & (70.70\ \text{Å}^3) & & (91.08\ \text{Å}^3) \end{array}\right\} \quad (23)$$

$$\text{Volume difference} = 0.88 \text{ per cent}.$$

Although there may be some question as to the propriety of calling this reaction "metasomatic," there is no question about the alternate second stages which usually follow that of equation (23). Two very opposite end-products may be obtained. Either the original bornite is almost entirely replaced by chalcocite, or it is almost completely changed to chalcopyrite. In contact with the average ore-forming fluid, the mixture of chalcopyrite and chalcocite does not appear to be a stable combination, perhaps because of the cupric and cuprous copper contained, respectively, in the two minerals. Of the two possible end-results, the more usual of attainment appears to be for the chalcopyrite formed to be replaced by chalcocite almost as soon as it is formed. Other conditions, however, appear to exist upon occasion which favor chalcopyrite as the more stable mineral, and it is formed not only at the expense of bornite but by the metasomatism of chalcocite as well. Edwards (1947, p. 104) discusses this reaction and gives two equations to explain the change of chalcopyrite to chalcocite. To condense the chemical expression of the reaction, these two equations have been combined so that they deal, after multiplying the first by 5, with comparable amounts of reactants and products; and the following modified equation results:

$$\left.\begin{array}{l} 20CuFeS_2 + 42Cu^{+2} + 10Fe^{+3} + 36H_2O \rightarrow 31Cu_2S + 30Fe^{+2} + 72H^{+1} + 9SO_4^{-2}\,, \\ (1414.00\ \text{Å}^3) \qquad\qquad\qquad\qquad\qquad (1411.74\ \text{Å}^3) \end{array}\right\} \quad (24)$$

$$\text{Volume difference} = 0.16 \text{ per cent}.$$

If equation (23) is multiplied by 20 and combined with equation (24), the result is equation (25), as follows:

$$\left.\begin{array}{l} 20Cu_5FeS_4 + 42Cu^{+2} + 10Fe^{+3} + 36H_2O \rightarrow 71Cu_2S + 30Fe^{+2} + 72H^{+1} + 9SO_4^{-2}\,, \\ (3264.20\ \text{Å}^3) \qquad\qquad\qquad\qquad\qquad (3233.34\ \text{Å}^3) \end{array}\right\} \quad (25)$$

$$\text{Volume difference} = 0.09 \text{ per cent}.$$

Edwards concludes that about 67.5 per cent of the copper in the chalcocite formed in equation (25) comes from the bornite and about 32.5 per cent is introduced from solution. From the figures given in the last three equations, the percentages are calculated as 70.42 and 29.58, respectively.

In equation (25) the oxidation of the extra sulfide ions in the bornite is indicated as being the companion reaction to the reduction of cupric to cuprous ion. A glance at table 2 will show that, under standard acid conditions, the second and third stages of the oxidation of sulfide ion to sulfate would not go forward because the oxidation potential of the cuprous-cupric half-reaction is greater than those of the two half-reactions of sulfide oxidation. If any close approximation of standard conditions exists in the ore-forming solutions, a more powerful oxidizing agent than cupric ion, e.g., ferric or manganic ions, would have to be present to drive the oxidation through the second and third stages. If, however, the ore-forming solutions are neutral or alkaline, the oxidation potential of the second and third stages of sulfide ion oxidation will be those for alkaline solutions and will be higher than that of the cuprous-cupric half-reaction, and the reaction probably will go as shown. This point is more fully discussed on page 529.

If we neglect the effectiveness of the oxidizing agent, the conversion of bornite to chalcocite is a relatively simple problem. That of changing bornite to chalcopyrite in more difficult. To maintain equality of volume, the 2 molecules of chalcocite formed in equation (23) must be converted to 1.3 molecules of chalcopyrite. This additional chalcocite is eliminated through the use of Stokes's equations as modified in equations (15*c*) and (16) of this paper. The result is equation (26):

$$\left.\begin{array}{l}\underset{(1411.74\ \text{Å}^3)}{31Cu_2S} + 62Fe^{+3} + 9S^{-2} \rightarrow \underset{(1414.00\ \text{Å}^3)}{20CuFeS_2} + 42Cu^{+2} + 42Fe^{+2}\,,\end{array}\right\} \quad (26)$$

$$\text{Volume difference} = 0.16 \text{ per cent}\,.$$

Again the ferric to ferrous iron reduction is appealed to as the agent to oxidize the cuprous ion to cupric, as was done in equation (22), for example.

THE REDUCTION OF FERRIC IRON AND ITS IMPLICATIONS

In formulating several of the equations above for hypogene replacement reactions, reduction of ferric to ferrous iron proved necessary to attain electrical balance between reactants and products. This is justified by analogy with the more completely known reactions of supergene enrichment and because of the almost exclusively ferrous nature of the iron found in sulfides and sulfosalts. The presence of many minerals containing substantial amounts of ferric iron in common igneous rocks would indicate a supply of ferric ions in the magma from which these rocks were derived. These ions, once the magma had cooled sufficiently to permit their formation, probably would be available for ejection from the magma chamber at any stage of the crystallization cycle by any mechanism adequate to the task. Granting that the ore-bearing fluid was formed from the magma at some point in that cycle (and before the supply of ferric ions was essentially exhausted, if that point ever is reached), it seems probable that such solutions would contain—along with much else, of course—ferric ion or complex ions of which ferric iron was a part. Yet ferric iron is rarely found in sulfide or sulfosalt minerals of ore deposits. Is it not reasonable, then, to assume that the process of hypogene ore deposition is, among other things, one in which the reduction of ferric to ferrous iron is of common occurrence?

SULFIDE REPLACEMENT OF CARBONATES

Because a common phenomenon in the formation of hypogene mineral deposits is the replacement of calcite or dolomite by ore sulfides, it is desirable to see if there is any difficulty in preserving volume and weight relations in the process. Equation (27) is a simple statement of the obvious initial components and final products of such a replacement:

$$Zn^{+2} + S^{-2} + CaCO_3 \rightarrow ZnS + Ca^{+2} + CO_3^{-2}. \tag{27}$$

This equation expresses merely a change of state and not a true chemical reaction, as the only difference between left- and right-hand components of the system is their transfer from solution to solid and vice versa. What causes the change of state is another matter on which no general agreement has been reached. The actual replacement is, in this case, apparently no more complicated than the diffusion of zinc and sulfide ions into the space initially occupied by the unit cells of calcite and the diffusion out of calcium and carbon and oxygen (or carbonate) ions. The role played by other ions present in the replacing solutions may have been vital in driving the reaction forward, but what those ions may have been and what forces they caused to operate is another matter. They have left no record of their presence in the solid minerals (unless it be in the content of liquid inclusions), and their part in the process is beyond the compass of this study to consider. Equation (27) represents a reaction which does take place, and that is the important thing so far as the balancing of that equation is concerned.

From table 1 it is obvious that equation (27) does not express a volume-for-volume reaction. It requires 1.55 formula unit volumes of sphalerite to replace 1 formula unit volume of calcite. This means that, getting rid of fractions, 31 formula units of sphalerite are required to replace 20 such units of calcite if a volumetric balance is to be achieved. Because the reactants in this equation do not have any ion in common, equation (27) is easily re-written as:

$$\left.\begin{matrix} 20CaCO_3 + 31Zn^{+2} + 31S^{-2} \rightarrow 31ZnS + 20Ca^{+2} + 20CO_3^{-2}, \\ (1220.60\ \text{Å}^3) \qquad\qquad (1220.47\ \text{Å}^3) \end{matrix}\right\} \tag{28}$$

$$\text{Volume difference} = 0.01 \text{ per cent}.$$

Because there is no chemical reaction here which requires balancing, there is no need to increase the number of components on either side of the equation beyond the requirements of volumetric equality. Were the compound replacing calcite of a larger unit cell volume than 122.06 Å³, then the number of metasome molecules involved would be less than the number of those of the host; but that is of minor significance. The important fact is that, in a replacement in which host and guest have no common element, molecular balance is not affected by the equating of unit cell volumes.

In this type of replacement the ions diffusing in and out of the crystal lattices are not solely the relatively small metallic ions but also include the quite large oxygen and sulfur ions as well. According to Fairbairn (1943, p, 1314) the sulfur ion (ionic radius = 1.74 Å) is about 1.5 times as large as the oxygen component of the carbonate radical. Its size, however, is small enough in comparison with the packing factors of the minerals involved[6] to permit the postulated diffusion.

NONMETALLIC REPLACEMENT OF CARBONATES

One possible nonmetallic replacement of carbonate is that of calcite by fluorite. It is a relatively simple reaction, shown in equation (29):

$$CaCO_3 + 2H^{+1} + 2F^{-1} \rightarrow CaF_2 + H^{+1} + HCO_3^{-1}\,, \tag{29}$$

which differs from those that have been taken up so far in that there is no change in the cation which is calcium before, after, and during the metasomatism. The equation is not, however, in volumetric balance because the unit cell of calcite is 122.06 Å³, while that of fluorite is 161.8. This leads to the conclusion that more calcium and fluorine must be added during the metasomatism than are required by molecular equality in the reaction. Equation (30) indicates one way in which this could be accomplished:

$$\left.\begin{array}{l} 2CaCO_3 + Ca^{+2} + 4H^{+1} + 6F^{-1} \rightarrow 3CaF_2 + 2H^{+1} + 2HCO_3^{-1}\,, \\ (122.06\ \text{Å}^3) \qquad\qquad\qquad (121.38\ \text{Å}^3) \end{array}\right\} \tag{30}$$

$$\text{Volume difference} = 0.56 \text{ per cent}\,.$$

Of course, equation (30) does not explain why calcium fluoride should be deposited from the ore-forming fluid in question or why the reaction of hydrofluoric acid on calcium carbonate should proceed at exactly or very nearly the same rate as does the deposition of fluorite. Nor does it explain why the action of hydrofluoric acid on calcite should be the removal of carbonate ion alone from the cell volumes concerned, without disturbing the calcium ions, which occupy so much smaller a portion of those volumes. An equation is not, however, supposed to do these things. It is an expression of the fact of chemical change, and beyond that it does not go. It would be of far greater value to be able to explain the reasons for the reactions that have been or will be discussed in this paper than to be able to write equations for them. Nevertheless, the explanation of the reactions may be somewhat facilitated, once it is shown that any such reaction can be expressed in a molecularly and volumetrically balanced equation properly indicating at least the correct initial and final products.

OXIDE REPLACEMENT OF SILICATES

If an attempt were to be made completely to balance molecularly and volumetrically the reaction taking place when a sulfide or an oxide replaces a silicate gangue mineral, it probably would fail because of the present lack of knowledge of the chemical characteristics of the ore-forming fluid concerned. Nevertheless, an equation can be written for the replacement, say, of wollastonite (whose formula is considered, for simplicity, to be $CaSiO_3$) by magnetite,

$$Fe^{+2} + 2Fe^{+3} + 4CaSiO_3 \rightarrow Fe_3O_4 + 4SiO_2 + 4Ca^{+2} \tag{31}$$

or

$$3Fe^{+2} + 6Fe^{+3} + 4CaSiO_3 \rightarrow 3Fe_3O_4 + 4Ca^{+2} + 4Si^{+4}\,, \tag{32}$$

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[6] Fairbairn gives a packing index of 4.0 for calcite. Because of the uncertainty of atom size in many sulfides, he gives no data as to the packing indices of sulfides. It appears that the index for sulfur itself lies between 6.5 and 7.0.

in either of the two above incomplete forms, from which much, of necessity, has been omitted.

In the molecularly balanced reaction expressed by equation (31), it is indicated that 260.00 Å³ of wollastonite are replaced by a total of 223.12 Å³ of magnetite and silica (73.40 Å³ from magnetite and 149.72 Å³ from silica). This leaves a deficit of 36.88 Å³, which cannot be supplied by silica, because the production of more silica can be attained only by the destruction of further wollastonite. Therefore, the 36.88 Å³ must be made up by the addition of magnetite (iron and oxygen both) from the solution. There is no more available oxygen to be derived from the wollastonite, once one molecule of magnetite has been formed from it, without breaking down further wollastonite, which would only further increase the volumetric unbalance of the system. Because 36.88 Å³ of magnetite equals 0.50 formula units of magnetite, it is obvious that $\frac{1}{2}$ formula unit of that mineral must be added to equation (31), in the ionic form on the left-hand side and in the solid state on the right. To eliminate fractions of a molecule, equation (31) must be multiplied by 2 and 1 additional molecule of magnetite must be added, as in equation (33), which follows:

$$\left.\begin{array}{l} 4O^{-2} + 3Fe^{+2} + 6Fe^{+3} + 8CaSiO_3 \rightarrow 3Fe_3O_4 + 8SiO_2 + 8Ca^{+2}\,, \\ \qquad\qquad (520.00\ \text{Å}^3)\quad (220.20\ \text{Å}^3)(299.44\ \text{Å}^3) \end{array}\right\} \quad (33)$$

Volume difference = 0.07 per cent.

In equation (32), 3 unit volumes of magnetite (220.20 Å³) appear to have replaced 260.00 Å³ of wollastonite. The deficiency in magnetite, is therefore, 0.54 formula unit volume, and this much must be added to the solid products from the solution. To eliminate fractions, it is necessary to multiply equation (32) by 20 and add 11 formula units of magnetite to the equation, in the ionic form on the left-hand side and in the solid form on the right, as shown in equation (34):

$$\left.\begin{array}{l} 71Fe^{+2} + 142Fe^{+3} + 80CaSiO_3 + 44O^{-2} \rightarrow 71Fe_3O_4 + 80Ca^{+2} + 80Si^{+4}\,, \\ \qquad\qquad (5200.00\ \text{Å}^3)\qquad\qquad (5211.40\ \text{Å}^3) \end{array}\right\} \quad (34)$$

Volume difference = 0.22 per cent.

These four equations—(31) through (34)— say nothing as to the manner in which the ferrous and ferric ions are carried in solution, nor do they supply any information as to the fate of the calcium and silicon ions (if the latter are produced at all) which are removed. It is almost certain that the silicon ion, if it forms, almost immediately becomes part of some acid radical, but the source of the other elements of the radical or what they may be is another problem.

It has been pointed out that the magnetite, above that provided by equations (31) and (32), must be present in the replacing solutions as ferrous, ferric, and oxygen ions or in ionic complexes from which these could easily be derived. The additional magnetite could not be formed by a later reaction of any other iron- and oxygen-bearing compounds because there would almost certainly be a change in volume from such a reaction which it would be simple to detect.

At the same time that the replacement reactions just discussed are taking place, one may also be going forward in immediately adjacent crystals of some other mineral in which oxygen ions must be diffused out. If this is so, a source of oxygen might be provided without recourse to the solutions carrying the iron ions. It would, however, be remarkable if the number of oxygen atoms not required in one reaction would just balance those needed in another, not to mention the difficulties of solid diffusion across grain boundaries. The need for the transfer of oxygen ions from the attacking solutions in some degree is then almost certain.

If the reaction of equation (33) actually has taken place on a large scale, many contact metamorphic magnetite ore bodies should be high in free quartz. This condition ordinarily does not obtain because the bulk of the iron oxide in such deposits appears to have been emplaced by the replacement of calcite rather than of contact silicates. If the reaction of equation (34) should occur, the formation of free quartz is unlikely; but the silicon ion produced must be removed, probably by the further silication of silicates already present in the immediate area, thus greatly complicating the reaction.

PACKING INDEX AND EASE OF REPLACEMENT

It seems reasonable, as suggested by Fairbairn (1943, p. 1351), that replacement of minerals having a high packing index (high ratio of ion volume to unit cell volume) should be more difficult to accomplish than that of those having a lower one. The replacement of calcite and wollastonite by magnetite or zincite bears this out. In the average contact deposit, the first stage of the process is the formation of a silicate tactite rock which, although high in typical lime-rich silicates, probably contains much recrystallized calcite. It is the consensus (Loughlin and Behre, 1933, p. 48; Schmitt, 1948, p. 4) that the oxides and sulfides which commonly are found in tactites are later than the silicates and possibly formed under less intense conditions. It is also commonly held that such oxides and sulfides usually replace the interstitial calcite in preference to the silicate portion of the rock. A possible explanation of this phenomenon is offered from Fairbairn's suggestion. Calcite has a packing index of 4.0, wollastonite one of 5.2. There is, therefore, over 20 per cent more "open space" in the calcite molecule through which ions can be diffused in and out of the lattice in calcite than in wollastonite. It would be only reasonable, then, in a rock containing a mixture of calcite and wollastonite (or other silicates and carbonates of similar packing relationships) to expect that replacement would be more likely to occur in the calcite.

OXIDE REPLACEMENT OF CARBONATES

The possible equations for such calcite replacements are of interest from the volume-weight standpoint as well. In equations (31) and (32) were given molecularly balanced equations for the replacement of wollastonite by magnetite. Similar equations can be written for a like replacement of calcite:

$$Fe^{+2} + 2Fe^{+3} + 2CaCO_3 \rightarrow Fe_3O_4 + 2Ca^{+2} + C + CO_2\,, \tag{35}$$

$$Fe^{+2} + 2Fe^{+3} + 4CaCO_3 \rightarrow Fe_3O_4 + 4Ca^{+2} + 4CO_2\,. \tag{36}$$

The theoretical basis for equation (35) is supplied by Winchell's work (1911, p. 229) on the formation of graphite. When the data of table 1 are used, these two equations can be volumetrically balanced, as follows:

$$\left.\begin{array}{l} 44O^{-2} + 31Fe^{+2} + 62Fe^{+3} + 40CaCO_3 \rightarrow 31Fe_3O_4 + 40Ca^{+2} + 20C + 20CO_2\,, \\ \qquad\qquad\qquad (2441.20\ \text{Å}^3) \quad (2275.40\ \text{Å}^3) \qquad (179.40\ \text{Å}^3) \end{array}\right\} (37)$$

Volume difference = 0.55 per cent ;

$$\left.\begin{array}{l} 28O^{-2} + 10Fe^{+2} + 20Fe^{+3} + 12CaCO_3 \rightarrow 10Fe_3O_4 + 12Ca^{+2} + 12CO_2\,, \\ \qquad\qquad\qquad (732.36\ \text{Å}^3) \quad (734.00\ \text{Å}^3) \end{array}\right\} \quad (38)$$

Volume difference = 0.22 per cent.

The main difference between equations (37) and (38) lies in the presence of graphite in the solid product of the former and its absence in that of the latter. In general it can be said that a comparative abundance of oxygen ion in the ore-forming fluid would favor the reaction of equation (37) and a deficiency in that ion, that of equation (38).

Two equations, analogous to (35) and (36), can be written for the replacement of calcite by zincite, as follows:

$$4Zn^{+2} + 2CaCO_3 \rightarrow 4ZnO + 2Ca^{+2} + C + CO_2\,, \qquad (39)$$

$$Zn^{+2} + CaCO_3 \rightarrow ZnO + Ca^{+2} + CO_2\,. \qquad (40)$$

These can be modified in the same manner as equations (35) and (36):

$$\left.\begin{array}{l} 19Zn^{+2} + 8CaCO_3 + 3O^{-2} \rightarrow 19ZnO + 8Ca^{+2} + 4C + 4CO_2\,, \\ \qquad (488.24\ \text{Å}^3) \qquad\qquad (447.64\ \text{Å}^3) \qquad (35.88\ \text{Å}^3) \end{array}\right\} \quad (41)$$

Volume difference = 0.97 per cent ;

$$\left.\begin{array}{l} 13Zn^{+2} + 5CaCO_3 + 8O^{-2} \rightarrow 13ZnO + 5CO_2 + 5Ca^{+2}\,, \\ \qquad (305.15\ \text{Å}^3) \qquad\qquad (306.28\ \text{Å}^3) \end{array}\right\} \quad (42)$$

Volume difference = 0.37 per cent.

Again the main difference between these last two equations, as was true of equations (37) and (38), is in the presence of graphite in the solid product of the one and its absence in the other. Both reactions require the addition of oxygen from the solutions; if the amount available is small, the formation of graphite will be favored. Reactions such as those of equations (37) and (41) may explain why graphite is found in pyrometasomatic deposits in pre-Cambrian limestones, such as that at Franklin, New Jersey.

Because of the relatively large proportion of oxygen in the calcite molecule, there would be an excess of oxygen in most ore oxide–calcite replacements, were it not for the necessity of removing some or all of the carbon of the calcite from the system. It is the need for oxygen to combine with the excess carbon, joined to the requirements of the ore oxide molecule, that forces the addition of oxygen from the ore fluids. As a result of this situation, oxygen (as free oxygen or oxygen ion) could be given up to the solutions only if the replacing ore oxide required so little oxygen as to leave enough to form CO_2 from all the carbon not changed to graphite and some to spare. If the oxygen so produced were to be converted into oxygen gas, the simultaneous reduction of some oxidizing agent, such as ferric ion, would be necessary.

CONCLUSIONS

The discussion presented in this paper indicates that it is possible to write equations for both supergene and hypogene replacement reactions which are balanced molecularly, volumetrically, and electrically. Complete equations for the average supergene metasomatism have been written which are probably close approximations of the truth. On the other hand, equations for hypogene reactions, in most cases, have been written only in skeleton form, though they are balanced in all respects.

To write the equation for a supergene replacement, the first step is to write the simplest molecularly balanced equation possible. This equation will be found to fit one of three general types. The first of these, type 1, shows that the volume of the replacing mineral provided is greater than the volume of the host. There must, therefore, be some additional amounts of the replaced mineral included on the left side of the equation and converted to soluble products on the right. This change is essentially one of oxidizing the sulfide ion to the sexivalent sulfur of the sulfate anion. This requires a strong oxidizing agent, and it is thought that the reduction of ferric to ferrous iron should be capable of doing the task. The second of these, type 2, shows that the volume of the replacing mineral produced by the reaction is less than that of the mineral replaced. It follows that more of the replacing mineral, both anion and cation, must be transferred from the liquid to the solid phase. Because the solubility of most sulfides, except hydrogen sulfide, in supergene solutions probably is negligible, the extra S^{-2} needed must come from H_2S or from the reduction of the sexivalent sulfur of the sulfate radical to the sulfide ion. This latter suggestion requires a strong reducing agent, and it is suggested that the oxidation of certain rather uncommon ions, such as Cd^{+1}, Mo^{+3}, etc., may drive the reaction forward. The third and last of these equations, type 3, indicates a volume of guest equal to that of host. Reactions for which equations of this type can be written appear to be exceedingly rare.

In this paper the main reliance, for both oxidation of S^{-} and reduction of S^{+6}, has been placed on half-reactions which undoubtedly occur in nature but which are not experimentally shown to do the work required of them in the volumetrically balanced equations here presented. It is pointed out that the oxidation of S^{-2} and the reduction of S^{+6} cannot take place in any area of chemical continuity at one and the same time.

The writing of equations for hypogene metasomatic reactions is far more difficult than expressing those which occur under supergene conditions. Too little is known of the composition and chemical nature of the ore-forming fluid to permit more than a skeleton presentation of the obvious essentials of the reaction. Again, however, three general types of equations can be set up. Type 1 includes those in which the amount of sulfur available in the host is greater than that needed to produce an equivalent volume of guest. This unwanted sulfur is removed by its oxidation to sexivalent positive sulfur of the sulfate ion. For this task the reduction of ferric to ferrous ion is considered adequate. Type 2 contains those reactions in which the amount of sulfur available in the host is less than that required in an equal volume of guest mineral. This additional sulfur is here considered to be supplied directly from the ore-forming solutions as sulfide ions. No opinion is expressed as to how these sulfide ions are made available by the solutions in question. It is possible that some of this sulfur may be produced by the reduction of sulfate ion, but it is believed that the evidence at hand is against a wide application of this method. Finally, in type 3 are placed those reactions in which sulfur in the host is exactly equal to that required by an equal volume of guest. Such reactions appear to be quite uncommon.

In hypogene reactions involving the replacement of carbonates and silicates by sulfides, fluorides, and oxides, three types of reactions also are recognized. If there is no common ion in such reactions, the problem of writing a balanced equation is a matter of equating volumes alone. This is true of replacements of oxides and silicates by minerals with sulfide or fluoride anions (or any of the many other anions not considered here). If both guest and host contain oxygen ions, however, the three possible types of reaction are similar to those just discussed for hypogene sulfide replacements. In all the equations of this type, oxygen is present in the calcite in amounts over and above the requirements of the ore oxide but is never in sufficient abundance to satisfy, as well, the needs of silicon or carbon which had to be converted to silica or carbon dioxide. Only in rare instances would the oxygen needed by the ore oxide be so low as to allow the release of oxygen as such to the solutions. If oxygen is deficient, it is shown as being added directly from the solutions as oxygen ions, and no attempt is made to explain its presence there in that form. No equations of the third, or originally balanced, type are given.

Finally, it must be remarked again that this paper is designed to show, not what reactions take place during hypogene and supergene replacements, but rather that equations for such reactions can be written and that they can be balanced molecularly, volumetrically, and electrically, once the data are available with which to write them. Those equations which have been presented are not the only possible ones that could have been written for the reactions with which they deal. They are, however, possible and in many cases even probable explanations of what occurs during replacement reactions. If these equations do nothing more, they demonstrate the truth of Emmons' remark (1940, p. 159) that replacement is definitely not a molecule-for-molecule exchange. It is much more than that, but that, too, is amenable to exact chemical expression.

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REVIEWS

"Die Metamorphose der Kohle und ihr Einfluss auf die sichtbaren derselben." By W. PETRASCHECK. (*Akad. Wiss., Math.-naturwiss. Kl., Sitzungsber. der österr.*, vol. 156, pt. 1, nos. 7 and 8.) Pp. 375–444.

This article is a study of coal metamorphism based partly on the published opinions of numerous individuals, on the author's extensive knowledge of European low-rank coals and coal beds, and on the results of experimentation with the autoclave, using wood and coals of various ranks, but mainly soft brown coal, in reproducing the effects of coalification artificially. This latter work was incidental to studies in the production of binderless briquets from lignite (soft brown coal). To those interested in the geological phenomena of coal metamorphism as an approach to the geochemical study of coal constitution this article will be of special interest.

At the start the author differentiates the process of coal formation from peat from the process of coal metamorphism. The first process is described as a biochemical process and receives little attention. The metamorphic process is a geological process and forms the main subject of consideration.

Metamorphism of coal is of two varieties—thermal and kinetic (p. 377), with differences often difficult to distinguish except immediately adjacent to igneous contacts. Metamorphism results in the so-called "coal series" of rank variation. In the low-rank coals (brown coal) Petrascheck lists the ranks of coal based upon differences in moisture content (p. 382) as shown in the accompanying tabulation.

Coal Type	Per Cent Moisture
Earthy brown coal	40–60
Soft brown coal	30–40
Matt-brown coal	20–30
Bright-brown coal	8–16

(American classification probably would draw the line between subbituminous coal and lignite at about 30 per cent moisture; the boundary between subbituminous coal and high volatile carbon bituminous coal possibly lies between matt-brown coal and bright-brown coal, but American differentiation is made on a B.t.u. basis, not on a moisture-content basis, so there is probably some overlapping.—G. H. C.)

What is called a "Schürmann's law" (p. 382) covers the relationship of moisture content to depth of cover for these low-rank coals and calls for a decrease of 1 per cent moisture for each additional 100 feet of cover, being based upon observations in East Borneo. Ordinarily low-rank coals do not have a depth of cover adequate to test the validity of the law.

Discontinuity of the coal metamorphic series is suggested (p. 383). (It may be indicated by the break in the moisture content between the matt-brown coal and the bright-brown coal as indicated in the table above. The author does not mention the work of Seyler in England, who claims discontinuity through the entire coal series on the basis of reflectivity studies.—G. H. C.) Petrascheck reserves opinion as to the discontinuity of the coal series in general.

Attention is called to the parallel consolidation effect of metamorphism upon coal bed and associated rocks, particularly clays and sandstones. Coal was pointed out by David White (p. 380) as a scale for measuring the milder stages of metamorphism.

The differentiation of peat and brown coal has been based upon the somewhat questionable distinction that moisture can be squeezed out of peat but not out of brown coal. Petrascheck cites evidence that it has been forced out of soft brown coal.

In general, it is the author's thesis (p. 388) that temperature above 100° C. enters very little into coal metamorphism except in thermal alteration. Static pressure is important up to a point, but tangential (kinetic) pressure is much more important. The unimportance of temperature is indicated particularly by the fact that resins, which melt at a relatively low temperature (less than 250° C.) have not been affected by metamorphism unless the coal has suffered thermal alteration.

The banded ingredients (vitrain, durain, etc.) are believed to exercise no influence upon coal rank.

The autoclave investigations (pp. 389–404) were carried on with small cubes of material cut from coal still containing the bed moisture. A pressure of 48–50 atm. was maintained in the autoclave, and a differential pressure, usually of 1 kg/cm,[2] was applied to the coal by weights at a temperature of 250° C. for 2 hours. The autoclave was always filled with water. Most tests were made with Köflach (Styria) soft brown coal. The coal was maintained in a moist condition to avoid changes inevitable if it was allowed to air-dry.

Volume and weight losses always appeared (p. 393). Some cubes were also deformed by unequal dimensional shrinkage. By these tests Petrascheck believed he demonstrated that metamorphism resulted in the destruction of cellulose, that lignified tissue produced vitrain, that fusain-like material was produced from green or dry wood, and that moor peat produced durain. Woods that have suffered much decay and have lost their cellulose before incoalation are much compressed.

That the autoclave tests were simply a mild form of low-temperature carbonization, as claimed by some (p. 395), Petrascheck denies, because there was not the characteristic production of tars and low-temperature gases, because there is selective disappearance of cellulose and the formation of a vitrain-like material from the wood, and because the pitchlike material that resulted resembled vitrain rather than tar, since it possessed typical wood structure in thin sections. Cell shrinkage corresponded to that observed in naturally formed coals.

In discussing the changes that take place in the structural components of coal (pp. 404–422), Petrascheck considers wood, fusain, durain, and the resins. The effect of metamorphism on wood is regarded as varied, depending upon whether cellulose had been lost through bacterial action before incoalation or later by metamorphism. In the first case the material is readily compressed with the loss of cell structure; in the latter case the presence of cellulose in the early stages of metamorphism tends to protect the cell structure, and the vitrain bands resulting from coalification are likely to preserve the cell structure, although the material consists largely of coalified lignin. (Undoubtedly, the claim that cellulose is lost through the metamorphic process will cause some lifting of the eyebrows on the part of botanists interested in the study of coal as a fossil material. Apparently, Petrascheck [p. 405] regards the good preservation of wood structure in vitrain bands as somewhat unusual. In Illinois coal beds the presence of such vitrain bands is of common occurrence, and bands have been seen as thick as 2–3 inches or more. The question may be raised whether the distinction between cellulose and lignin is sufficiently precise so that the layman can be sure that there is general agreement among botanists concerning the meaning of the terms.—G. H. C.) Petrascheck calls attention to the fact that, under the particular P/T conditions applied, cellulose disappears in the autoclave. It also disappears, he says, between matt and glance brown-coal stages of metamorphism (p. 406). (Presumably that means that fossilized cell structures representing cellulose are not present in coals of bituminous and higher ranks, a conclusion that the American coal paleobotanist will probably not readily accept without substantial proof.—G. H. C.)

Fusain is discussed at length (pp. 408–415) and, although the objections to the forest-fire explanation are admitted—viz., the presence of resin in fusain, the rank of fusain, and the manner of occurrence of fusain—he concludes that the forest fire is the most probable mode of origin of most fusain. He has definite reservations in regard to the possible importance of hydrogen-ion concentration in the associated strata.

Petrascheck believes that evidence of "burned layers" is more common than has generally been conceded and that their absence in some coal beds is due to the fact that such beds were "low-moor" deposits periodically overwhelmed by water. Many other beds not so situated do contain "charcoal" layers.

The Mackenzie-Taylor theory of base exchange is interesting but inadequately proved. The author admits, however, that kaolinized bed partings in coal beds are commonly associated with abundant fusain, whereas fusain is not common in beds in which montmorillinite forms the chief mineral in clay partings.

The idea that fusain has a more or less fixed rank does not seem to be borne out by the evidence, according to Petrascheck. Using two charts (pp. 410, 412), in which volatile matter of fusain and of whole coal or of vitrain is plotted against rank, he shows that the variations in the fusain roughly parallel those in the whole coal or vitrain from the same beds, volatile matter decreasing with the rank of the coal and fusain.

The occurrence of resins in fusain (p. 413) is not regarded as fatal to the forest-fire theory, mainly, it appears, because charcoal may yield bituminous substances. Possibly the resin problem is dismissed too lightly.

Various other considerations in regard to fusain are raised and explained, such as the fusainization of the inner or outer portions of stems, as the case may be, and the presence of fusain in coal balls, indicating that it originated in the moor deposit. The conclusion is reached that the forest-fire explanation is preferred to any other that has been proposed.

Durain (p. 416) is regarded by Petrascheck as coalified moor coal. The humic material present is derived from leaf debris rather than from woody stems. It is generally but not always noncoking, since the "coking capacity" of coal resides mainly in vitrain. Volatile matter is somewhat higher in durain than in vitrain, with which it may be associated. The origin of the opaque matter in durain is said to be the humic acids, but the explanation is unsatisfactory (p. 417). (No reference is made to explanations of opaque matter that have been advanced by R. Thiessen from time to time.—G. H. C.) Durain is regarded as a quiet-water deposit by Petrascheck and the association of durain with sponge spicules, diatoms, and other aquatic forms is cited as evidence. (The durain of Petrascheck is evidently both the clarain and the durain of British and American writers, as well as the opaque and translucent attritus of Thiessen. There is less discrimination with respect to the varieties of "durain" than seems to be employed by coal petrographers in this country.—G. H. C.)

Resins (pp. 418–422) of two kinds are recognized: resins representing plant secretions, called "retinite" or "resinite," and resins which are designated as "exudations." This latter material fills cracks and lies along bedding planes and apparently includes the so-called "organic minerals" known as duxite, hartite, and köflachite. It is relatively rare, and whether or not such material is found in American coals is not known by the reviewer.

Pollen, spores, cuticles, and waxy material provide the source of what Petrascheck calls the "polymer bitumens." The spores have important uses in stratigraphic geology. In general, spores and pollen in low-rank coals have the chemical characteristics of modern pollen. They provide a means of checking the temperature to which coals have been subjected, since they melt at less than 200° C. and decompose at temperatures between 300° and 350° C. In general, these substances disappear in the high-rank coals.

Finally, there is the material called "bituminous wood" by Petrascheck (p. 424), being a coal of a special light yellow color, displaying cellular structure, the megascopic appearance of vitrain, but very "bituminous." It is believed by the author of the paper to have originated from decayed wood.

One section of the article is given to the consideration of the effect of the hydrogen-ion concentration in the coal beds and in associated strata. High moors and forest peats are said to have an acid environment; low moors a neutral or weakly acid environment. Kaolin forms in an acid medium and montmorillinite in an alkaline medium. In general, as has been previously noted, fusain seems to be more abundant in coal beds containing kaolinitic clays than in those in which the clay is largely montmorillinite. (The extent to which these generalizations may apply to American coals is not known.— G. H. C.)

The latter part of the paper (pp. 429–439) concerns the tectonic evidences of metamorphism. These consist mainly in the development of joints. There is general increase in the amount of jointing with the degree of metamorphism, and jointing is more characteristic of tectonic than of thermal metamorphism. Optical changes (pp. 431–432) accompany alteration, the low-rank coals being isotropic and the higher-rank anisotropic. The boundary lies between the matt-brown coal and the glance or bright-brown coal. The extinction tends to be parallel to the bedding of a coal bed when the latter is horizontal or vertical; otherwise it is inclined to the bedding.

Special studies were made of the coals near Handlova, Czechoslovakia, Pernik, Bulgaria, and Salgotarjan, Hungary, to determine the relative importance of tectonic and thermal metamorphism. The discussion has local significance only.

The metamorphism of Sudetenland, Germany, brown coal (pp. 432–439) is also discussed, tectonic versus thermal metamorphism also being at issue. Jurasky has spoken for thermal metamorphism from a deep-lying magmatic center. Petrascheck, speaking for tectonic metamorphism, believes that gas bubbles in the resin, bitumen exudations (duxite), and a peculiar type of columnar cleavage in the coal, all of which are cited by Jurasky as evidences of thermal alteration, are found in coal beds that have been permanently cool. Other evidence is cited for doubting even the existence of the deep-seated igneous center.

Toward the end of the article (p. 439) the author calls attention to the possible usefulness of kaolin and bentonite as geological thermometers. The first loses moisture at 450° C., and bentonite at 250° C. The evidence supplied by these minerals indicates that the Sudetenland coals did not reach the high temperatures suggested by Jurasky.

In summarizing the summary (pp. 439–442) note may be made of the following conclusions. Absolute differentiation of thermal and kinetic metamorphism is impossible. Volatile-matter content is not a suitable basis for the classification of brown coal; moisture content is satisfactory. Depth of overburden is not adequate as a sole cause of rank advance. Wood (vitrain) is the most reactive substance in coal, but cellulose is lost in the coalification process. Rotten and cellulose-free wood compresses very flat but possesses the megascopic appearance of vitrain. Moor coal is the source of durain. Autoclave tests substantiated the loss of cellulose under pressure. Fusain is generally derived from wood charcoal. Undulatory extinction is an attribute of resins of high-rank coal. The exudation of resin does not necessarily require thermal metamorphism. The degree of jointing and its character are a measure of the tectonic stress, as is likewise anisotropism beginning with bright-brown coal.

The article is well documented with 111 references mainly to German and Austrian literature but with a few English and American citations. Unfortunately, the references are not always adequately described so that they can be easily found.

G. H. Cady

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PETROGENESIS OF PERTHITE PEGMATITES IN THE BLACK HILLS, SOUTH DAKOTA[1]

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ABSTRACT

The study of some pegmatites of the Black Hills, South Dakota, has led to the conclusion that metasomatic replacement played a significant role in their formation. Microcline microperthites of these pegmatites were not products of magmatic crystallization but were formed by complex processes of perthitization. The hypothesis of the metasomatic origin of the perthites is supported by chemical analyses and microscopic features of the perthites and country rocks and by field observations.

STATEMENT OF PROBLEM

The origin of the potash-rich rocks, whether deep seated or extrusive, is still an unsolved problem. Some authors explain them as products of magmatic crystallization, whereas others believe in a secondary mode of origin of the potash. Final conclusions regarding the genesis of these rocks can be drawn only on the basis of field and laboratory investigations of individual occurrences.

The author studied some potash-rich pegmatites in T. 2 S., R. 6 E., near Glendale, Pennington County, South Dakota (fig. 1). Excellent outcrops and mine cuts facilitated the study. The pegmatite mines studied include the Glendale, in the northernmost portion of NW. ¼ Sec. 23, T. 2 S., R. 6 E.; the Soda Spar, in the SE. ¼ Sec. 22, T. 2 S., R. 6 E.; the Johnson Dike, in the SE. ¼ NW. ¼ Sec. 22, T. 2 S., R. 6 E.; and several smaller unnamed bodies. Several pegmatites, such as the Three Falls Lode in the SW. ¼ Sec. 22, T. 2 S., R. 6 E., previously mined, were also examined.

Smaller pegmatites are very numerous in this area. The map of pegmatites of the Beecher Rock Basin, T. 4 S., R. 4 E., in Custer County, South Dakota, made by Gwynne (1944) shows clearly their wide distribution.

Some of the largest pegmatites of the Black Hills in the neighborhood of both Custer and Keystone were also visited. These were the Buster, 2½ miles south-southwest of Custer, in the SW. ¼ Sec. 2, T. 4 S., R. 4 E.; the New York, in the NE. ¼ Sec. 18, T. 4 S., R. 4 E.; the Peerless, in the NE. ¼ SE. ¼ Sec. 8, T. 2 S., R. 6 E.; the Hugo, in the southernmost portion of SE. ¼ Sec. 8, T. 2 S., R. 6 E.; the White Cap, in the S. ½ Sec. 9, T. 2 S., R. 6 E.; and the Etta, in the NW. ¼ NW. ¼ Sec. 16, T. 2 S., R. 6 E.

The structure of most of these pegma-

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[1] Manuscript received October 8, 1948.

[2] On a study leave at the University of Chicago.

tites has been described by other writers: Apsouri (1940), Connolly and O'Harra (1929), Guiteras (1940), Fisher (1942 and 1945), Landes (1928), Lincoln (1927), Page and Norton (1946), and Tullis (1939). Few chemical or microscopic data have been published. The author, therefore, analyzed some perthites and country rocks and examined 65 slides of these rock types. Still more data are needed to make certain the conclusions drawn.

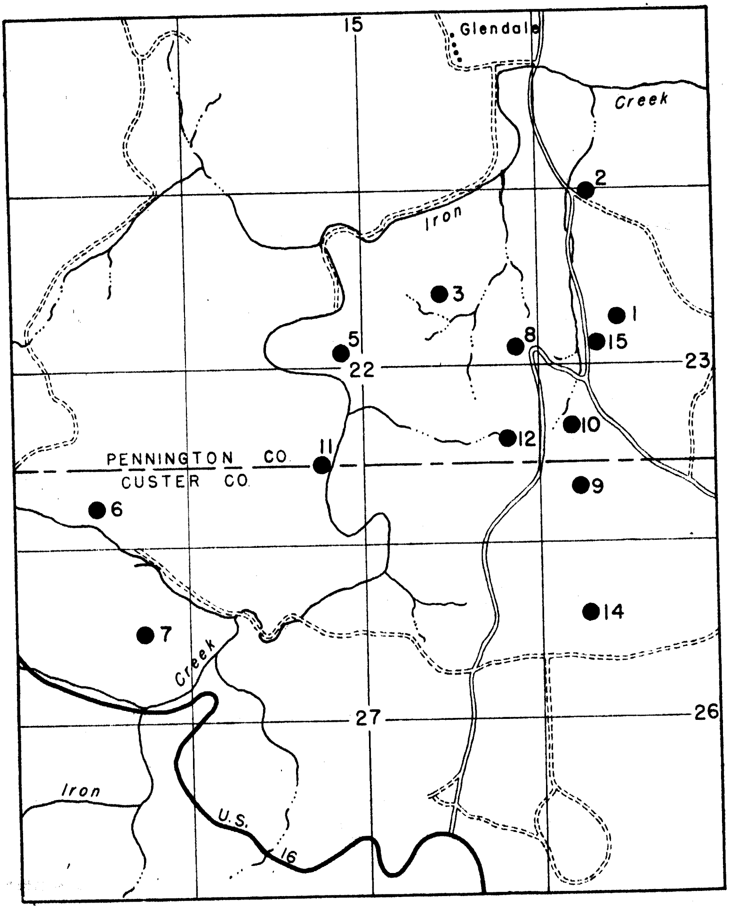

Fig. 1.—Glendale area, showing location of perthites

GEOLOGY OF PEGMATITE AREA

The general geology of the Black Hills has been discussed by Darton and Paige (1925). The Black Hills uplift, an outlier of the Rocky Mountains, is an elliptical area whose long axis trends northwest. The core is composed of metamorphic and eruptive rocks surrounded by concentric outcrops of Paleozoic and Mesozoic sediments. The pre-Cambrian central portion and the outwardly dipping sediments are overlain by Oligocene formations. Hypabyssal dikes and sills of Tertiary age intrude the pre-Cambrian metamorphic rocks and the Paleozoic and Mesozoic sediments (Kirby, 1932).

The pre-Cambrian rocks are metamorphosed sediments and basic igneous rocks, granites, and granite pegmatites. The sediments are conglomerate, quartzite, slate, phyllite, mica schist, garnet and staurolite schists, and marbles (Van Hise, 1890). The basic rocks, originally dioritic or gabbroic, are amphibolites and hornblende schists. In places metamorphism makes it difficult to distinguish between metabasics and metasediments (Connolly, 1927).

The pegmatite granite of Harney Peak is a coarse-grained potassic granite composed principally of microcline-perthite, quartz, and muscovite with an unusually large amount of tourmaline. The pegmatites are also rich in potash and consist chiefly of those minerals which form the pegmatitic granite.

Intense metamorphism and lack of fossils (Connolly, 1927) make it difficult to determine the age of the sediments. Most investigators assign these rocks to the Algonkian. Davis (1926) determined the age of the uraninite from the pegmatites of the southern Black Hills as 1,167 million years. The corrected age, applying Wickman's graphic chart (1943) of $U/(U + Pb)$ and $U/(U + Th)$, is 1,450 million years. The Black Hills pegmatites are, therefore, Archean.

FIELD RELATION OF PEGMATITE AND ASSOCIATED SCHISTS

Recent studies of the Black Hills pegmatites show that most of the mined pegmatites are zoned (Page and Norton, 1946). The zones occur in a definite sequence, and, although any number of different zones may be missing, the sequence of the existing zones will remain the same. Perthite may be either an essential constituent of the zones or may form one of the zones. The Hugo pegmatite and the High Climb Pit, 6 miles north of Custer (Fisher, 1942, pl. 19 and fig. 4) are examples of the first type. Perthite is the principal constituent of some of the different zones of the Soda Spar pegmatite, the Johnson Dike, and the Three Falls Lode.[3]

Examples in which perthite alone forms one of the zones are the White Cap and the John Ross Lode, 4 miles west of Custer (Fisher, 1945, pp. 80–82). The Glendale pegmatite consists of perthite only. Fisher (1945, p. 10) has shown that pegmatitic masses composed almost entirely of perthite are common.

Abundant pegmatitic pockets in numerous unzoned bodies in many cases are composed almost entirely of perthite, commonly associated with albitites.

Perthite is also found as metacrysts in the perthitized schists and as coarse phenocrysts in the granites of the area. The unaltered schist is composed essentially of muscovite, biotite, and quartz with minor amounts of garnet. The schist in places is feldspathized and/or tourmalinized.

The feldspathized schists contain from 25 to 80 per cent of feldspar. The feldspar may be a perthite, an albite, or a microcline.

An outcrop of the perthite schist (pl. 1, *A*, *D*) about 36 yards long, lies in the NE. ¼ Sec. 28, T. 2 S., R. 6 E. The foliation of the perthitic schist, of the schist inclusions commonly found in the pegmatites, and of the isolated schist outcrops all have the same trend. The perthite of the schist appears mainly along the schistosity as small ovoid masses less than 1 mm. in diameter, grading into masses more than 3 cm. long, which tend to assume the rectangular outlines of a true feldspar (pl. 1, *A*). Some of the larger metacrysts have biotite laminae curved around them, indicating that the growing feldspar crowded aside the folia of the schist. The ovoid masses sometimes join and form a pinch-and-swell structure. Perthites crowded with many megascopic schist inclusions grade into uncontaminated perthite (pl. 1, *B* and *C*). Replacement rather than viscous intrusion and magmatic crystallization seems to be the best explanation of these phenomena.

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[3] For location see p. 555 and fig. 1.

Microscopic features support this view. The inclusions in the perthite metacrysts are brown biotite and quartz. The biotite has the same optical properties as the schist. The quartz inclusions are more or less rounded and elongated in a direction parallel to that of the schistosity of the matrix. Some of them still

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PLATE 1

A, Perthite of the perthite schist assuming the rectangular outline of the feldspars (white with dark schist inclusions). Pushing-aside of the schist foliae is best shown at the corners of the crystal *P*.

B, Perthite, uncontaminated crystal (*white*); another smaller crystal appears to the right of the figure. The groundmass is schist relic, showing black quartz and lighter-colored micas.

C, A crystal of perthite of the perthite schist, showing inclusions of the schist in a megascopic sieve texture. Parts of the crystal are uncontaminated (*white*). Inclusions are dark.

D, Perthite schist, showing many perthite crystals (white with light-gray schist inclusions). Schist relic is dark and forms the groundmass.

PLATE 2

A, Perthitic schist showing albite (*A*) and string perthite (*P*) with faint grating. Note (1) rounded quartz inclusions (schist relic) in both albite and perthite, (2) the cementing material between some quartz grains is still preserved (*C*), (3) the trend of the quartz inclusions is roughly northeast-southwest in the figure, and (4) biotite inclusions (*B*). 12×.

B, Feldspathized schist showing albite (*A*)—polysynthetic twinning—and potash feldspar (*PF*)—simple twinning, Carlsbad law. Notice (1) rounded inclusions of quartz in the feldspars (in some feldspar crystals the inclusions appear white, in others they are extinguished); (2) the trend of inclusions is the same in the different feldspar crystals—northeast-southwest; and (3) the pushing-aside of the biotite flakes of the original schist (*B*) at the corners of the feldspars. 12×.

C, Albitite. Notice the cementing material (*C*) between some quartz grains. Quartz (*Q*), albite (*A*), tourmaline (*T*). The albite crystals show polysynthetic twinning, all trending in a roughly parallel direction. Most of them do not show up clearly in the figure. 12×.

D, Section of a perthite parallel to (001), showing patch (*white*) and microcline (*grated*). Note the microcline island in the albite patch in optical continuity with host microcline. Note also the extinguished ungrated portion of the microcline. Albite and microcline edges show embayments. 41×.

E, Section of a perthite parallel to (010), showing albite veins and patches (*white*) with numerous microcline islands (*gray*). 12×.

F, Section of a perthite parallel to (001), showing patch albite (polysynthetic twinning) and microcline (*grated*). Note microcline relics in the patch, showing optical continuity with the host microcline, and irregular resorbed contacts. 12×.

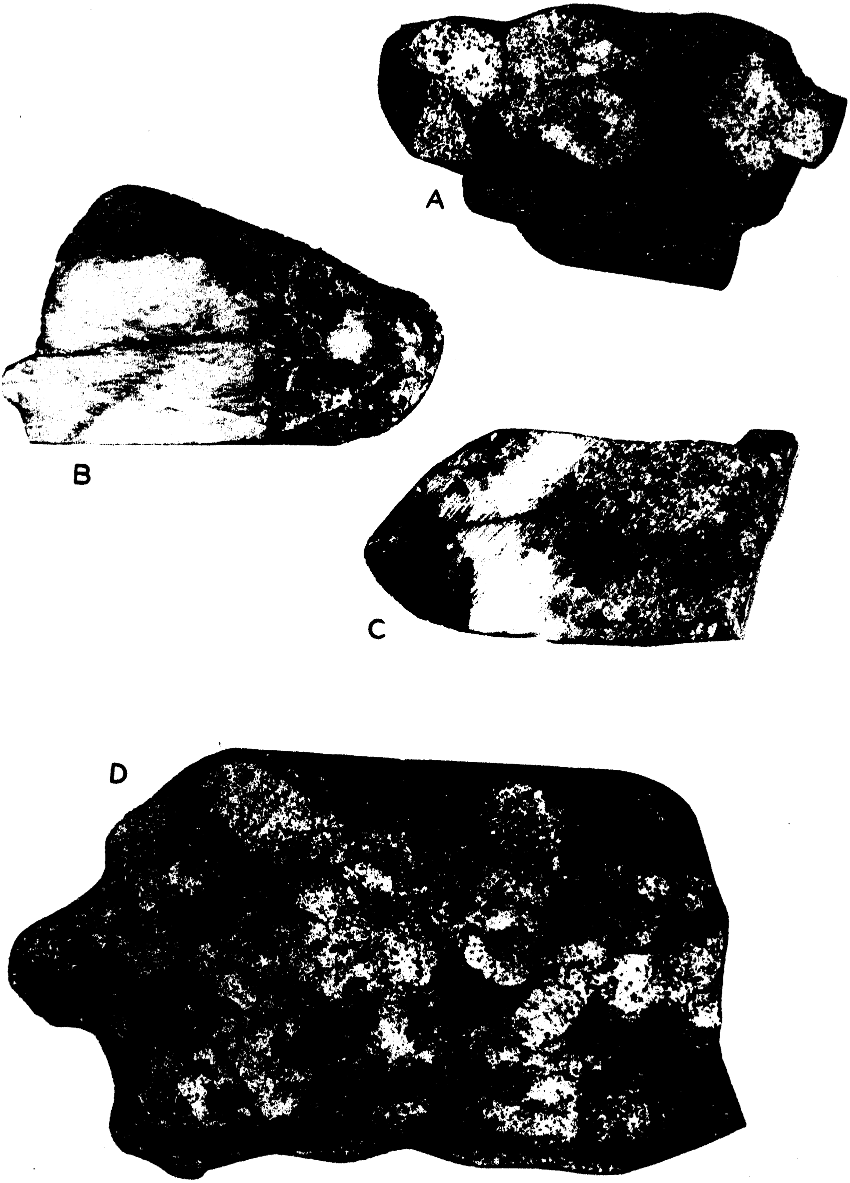
PLATE 3

A, Patch and vein perthite (no. 13, table 1), section parallel to (010), albite (*white*), microcline (*gray*). Note (1) fine strings and thin films between the veins, (2) grading of the strings and films into veins near the center, (3) grading of the veins into patches, and (4) microcline relics in the patch at the lower portion. 12×.

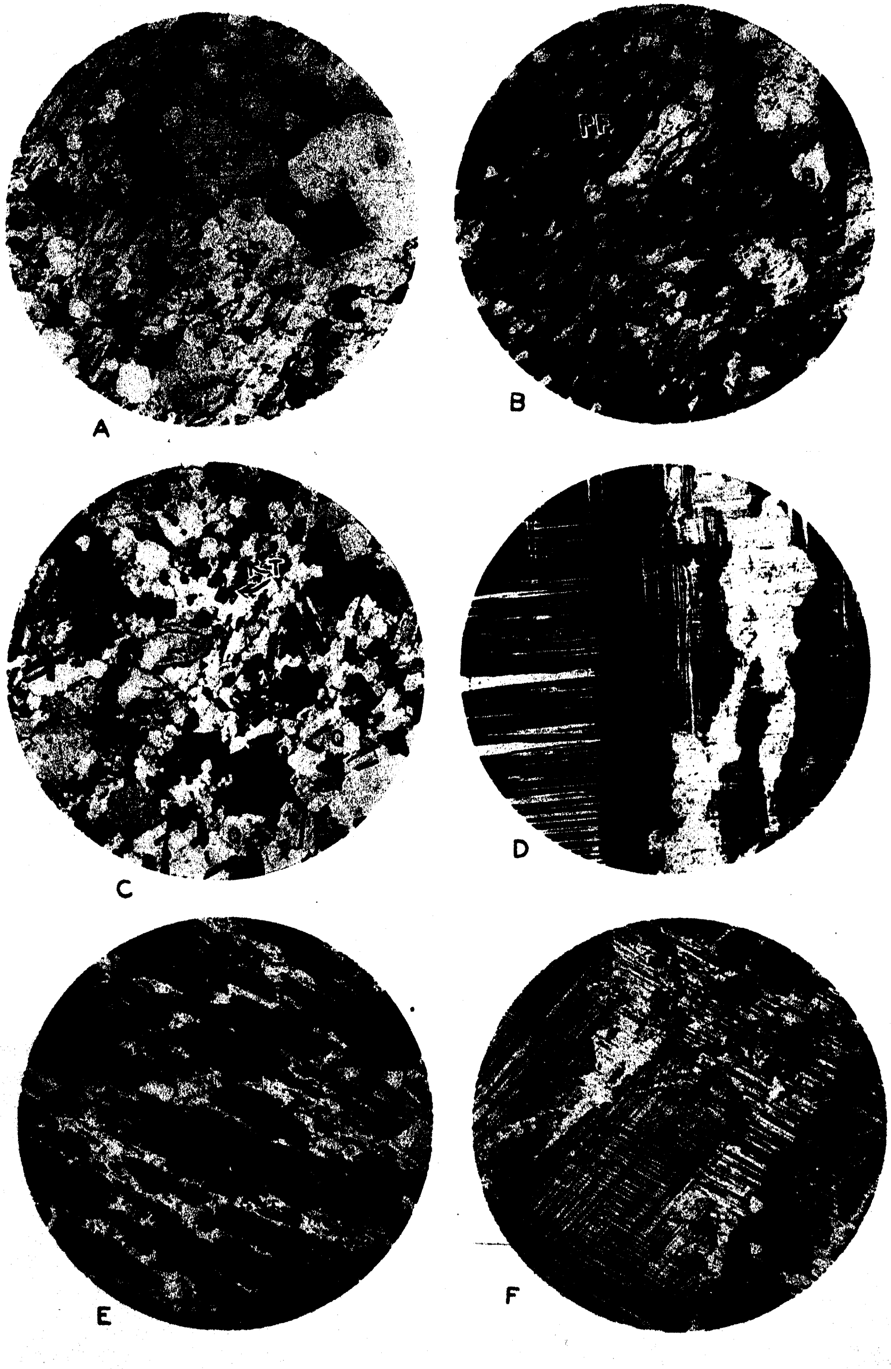
B, Film, vein, and patch perthite (no. 5, table 1), section parallel to (010), albite (*white*), microcline (*dark gray and black*). Note (1) abundance of thin films, (2) grading of the films into veins, (3) complex texture of the veins grading into patches, and (4) microcline islands in the patch near the center. 12×.

C, Film and vein perthite (no. 7, table 1), section parallel to (001), albite (*black*), microcline (*finely grated*). Note abundance of regularly elongated veins and thin films. The latter grade into the former. 12×.

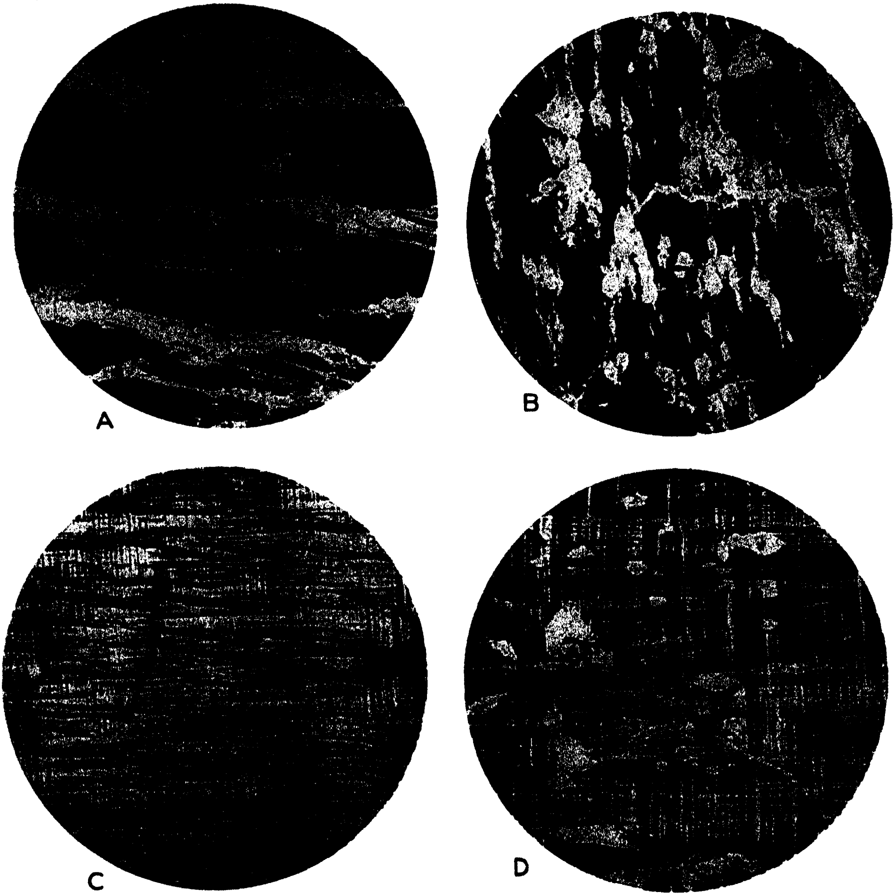
D, Vein perthite (no. 10, table 1), section parallel to (010), albite (*white*), microcline (*grated*). Note the roughly round cross section of veins. 12×.



Perthite of the perthitized schist



Thin sections of feldspathized schist and perthite



Thin sections of perthite

have the yellowish-green chloritic or the yellowish-brown ferriferous cementing material, like that of the quartz of the original unaltered schist (pl. 2, *A*). The albite of the schist has formed in the same manner as the perthite. Plate 2, *A* shows inclusions of rounded quartz of the schist in the albite. They are elongated in the direction of the schistosity. Albite from a schist in the NE. ¼ Sec. 8, T. 2 S., R. 6 E. (across the east-west road from the mill of Consolidated Feldspar Corporation near Keystone) contains abundant quartz inclusions which clearly show a sieve texture (pl. 2, *B*). All the inclusions in the feldspar crystals are elongated in a northeast-southwest direction in agreement with that of the schistosity and the quartz of the matrix. The albite schist in some cases grades into albitites composed essentially of albite, quartz, and micas. In these albitites the cementing quartz material can still be observed between some quartz grains, and the albite is generally elongated in the direction of the schistosity (pl. 2, *C*). These transitional rocks, containing some relic features of the schist, suggest a metasomatic origin for both the perthite schists and the albitites.

COMPOSITION OF PERTHITES AND RELATED ROCKS

Fifteen perthites, one cleavelandite, and three samples of the country rocks were analyzed chemically by the writer at the University of Chicago. Specimens of perthite and microcline from the Hugo pegmatite, obtained from Dr. L. Page, of the U.S. Geological Survey, were analyzed by Mr. B. Bruun. The analyses of the sixteen perthites and the microcline appear in table 1, the analysis of the cleavelandite is given in table 2, and the analyses of the country rocks are shown in table 3.

The normative potassium feldspar (Or), sodium feldspar (Ab), and lime feldspar (An) of each of the analyzed perthites, microcline, and cleavelandite were calculated, using the equivalent molecular percentage instead of the molecular percentage (Niggli, 1936). This is shown in table 4. The Or is the molecule $KAlSi_3O_8$, composed of one atom of potassium, one of aluminum, and three of silicon; the Ab is the molecule $NaAlSi_3O_8$; and the An is $CaAl_2Si_2O_8$. The percentages of the three feldspar molecules do not differ much from those obtained with the weight percentages in calculat-

TABLE 1

ANALYSES OF PERTHITES*

Oxides	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	Average	16	17
SiO_2	63.89	65.13	64.29	63.67	64.48	64.36	63.96	64.13	64.11	64.30	64.29	64.48	65.10	64.97	65.07	64.41	64.69	64.20
TiO_2	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al_2O_3	21.61	20.70	19.40	20.08	19.54	19.53	19.39	18.92	19.62	19.37	19.76	19.66	19.55	19.41	19.04	19.72	19.79	19.38
Fe_2O_3	0.09	0.09	0.09	0.07	0.09	0.09	0.13	0.06	0.06	0.09	0.09	0.12	0.10	0.07	0.13	0.09	0.17	0.05
MgO	0.00	0.03	0.11	0.02	0.00	0.05	0.05	0.07	0.12	0.03	0.06	0.06	0.00	0.02	0.03	0.04	0.01	0.02
CaO	0.78	0.25	0.32	0.49	0.79	0.82	0.46	0.15	0.37	0.27	0.27	0.21	0.07	0.36	0.17	0.38	0.02	0.02
BaO	0.02	0.05	0.06	0.02	0.01	0.02	0.17	0.03	0.04	0.05	0.03	0.05	0.02	0.05	0.05	0.04	n.d.	n.d.
Na_2O	2.72	2.94	0.38	1.96	2.74	2.96	2.80	2.12	2.55	1.43	2.89	2.12	2.98	3.06	3.29	2.46	3.19	1.21
K_2O	11.08	10.58	15.62	13.68	12.52	11.94	12.41	14.03	12.84	14.02	12.30	13.00	11.58	12.35	11.94	12.66	11.95	14.95
H_2O	0.31	0.17	0.16	0.18	0.21	0.29	0.36	0.25	0.22	0.23	0.21	0.28	0.16	0.12	0.16	0.22	0.27	0.27
Total	100.50	99.94	100.43	100.17	100.38	100.06	99.73	99.76	99.93	99.79	99.90	99.98	99.56	100.31	99.88	100.02	100.09	100.10

* 1–15: Perthites (see fig. 1 for location); analyst, R. A. Higazy. 16: Perthite (Hugo pegmatite); analyst, B. Bruun. 17: Microcline (Hugo pegmatite); analyst, B. Bruun.

ing the norm. The sum of the three molecules forms in most cases 98 per cent of the total normative composition.

NORMATIVE COMPOSITION OF THE PERTHITES

The normative ratios Or:Ab and Ab:An of the feldspars shown in table 4 were plotted in figure 2. Perthite 1 has an Or:Ab ratio of 72.5:27.5, similar to that of perthite 13, which has the ratio 72.0:28.0. Yet the Ab:An is 88.5:11.5 in the former and 98.5:1.5 in the latter. The similarity of the Or:Ab ratio in the two perthites might indicate that their temperatures of formation were about the same, and, accordingly, one would expect them to have similar normative plagioclase composition. On the contrary, the normative plagioclase in the former is near oligoclase, whereas in the latter it is sodic albite. Similarly, perthite 14 and 6 have Or:Ab ratios of 73.1:26.9 and 73.2:26.8, respectively, and Ab:An ratios of 95.5:3.5 and 87.7:12.3, respectively. Moreover, figure 2 shows that there can exist perthites with similar normative plagioclase composition, but different Or:Ab ratios. Examples of this are: (1) perthites 15 and 8, which have Ab:An ratios of 97.0:3.0 and 96.5:3.5, respectively, and Or:Ab ratios of 70.7:29.3 and 81.3:18.7, respectively; (2) perthites 4, 5, and 6 with Ab:An ratios of 88.0:12.0, 87.4:12.6, and 87.7:12.3, respectively, and Or:Ab ratios of 82.0:18.0, 75.2:24.8, and 73.2:26.8, respectively.

TABLE 2

ANALYSIS OF CLEAVELANDITE*

Oxides	Per Cent Oxides
SiO_2	67.62
TiO_2	0.00
Al_2O_3	19.76
Fe_2O_3	0.09
MgO	0.08
CaO	0.50
BaO	0.05
Na_2O	11.03
K_2O	0.09
H_2O	0.40
Total	99.62

* Analyst, R. A. Higazy.

Perthite 13, found as a phenocryst in the granites of the area, has a normative plagioclase composition of Ab 98.5. This is more sodic than most of the perthites. If this phenocryst had crystallized from a granitic magma, it would have a normative plagioclase poorer in the albite molecule than that of the pegmatite perthites because it formed at a comparatively higher temperature. Moreover, the normative plagioclase composition of cleavelandite (No. 18, table 4) is Ab 97.5 An 2.5. This is richer in anorthite than is the perthite phenocryst of the granite. Also, perthite 2 has exactly the same normative plagioclase composition as that of cleavelandite. Perthite 13 has even a lower anorthite content than does cleavelandite. Therefore, it can be stated that such perthites might have formed at the same temperature as cleavelandite (the

TABLE 3

CHEMICAL ANALYSES OF THE COUNTRY ROCKS*

Oxides	1	2	3
SiO_2	68.83	70.85	69.93
TiO_2	0.46	0.21	0.00
Al_2O_3	14.63	15.48	17.84
Fe_2O_3	1.42	0.64	0.14
FeO	3.83	1.09	0.20
MnO	0.05	0.01	0.01
MgO	2.24	0.83	0.54
CaO	0.52	0.38	0.59
BaO	0.11	0.11	0.02
Na_2O	1.05	2.16	9.68
K_2O	5.01	6.45	0.42
P_2O_5	0.17	0.45	0.51
H_2O^+	1.52	1.18	0.23
H_2O^-	0.14	0.08	0.06
Total	99.98	99.90	100.17

* 1, Biotite schist; 2, microcline-perthite biotite schist; 3, albitite. Analyst, R. A. Higazy.

hydrothermal species), if not lower. Lastly, 17 is a microcline collected from the microcline-quartz zone of the Hugo pegmatite, and 16 is a perthite from the perthite zone. The perthite which is from an outer zone and supposed to have crystallized at an earlier stage than the microcline of the inner zone has a normative plagioclase composition of Ab 99.7 An 0.3; this is more sodic than that of the microcline, which has the normative plagioclase Ab 99.0 An 1.0.

COMPOSITION OF PERTHITE OF THE SCHIST

The similarity in composition of the perthite metacryst in the schist and the phenocryst in the granite with the perthites of the pegmatites is striking. Table 5 shows that the weight percentages of all the oxides of a perthite meta-

TABLE 4

RATIO OR: AB: AN OF THE DIFFERENT PERTHITES*

Number	Or	Ab	An	Ab: An	Per Cent Plag.	Or: Ab
1	70.4	26.2	3.4	88.5:11.5	29.6	72.5:27.5
2	69.4	29.3	1.3	97.5: 2.5	30.6	70.2:29.8
3	94.7	3.5	1.8	66.0:34.0	5.3	96.2: 3.8
4	80.1	17.5	2.4	88.0:12.0	19.9	82.0:18.0
5	72.5	24.0	3.5	87.4:12.6	27.5	75.2:24.8
6	70.5	25.9	3.6	87.7:12.3	29.5	73.2:26.8
7	72.8	24.9	2.3	91.6: 8.4	27.2	74.5:25.5
8	80.7	18.6	0.7	96.5: 3.5	19.3	81.3:18.7
9	75.5	22.7	1.8	92.7: 7.3	24.5	76.8:23.2
10	85.3	13.3	1.4	90.6: 9.4	14.7	86.5:13.5
11	72.7	25.9	1.4	95.0: 5.0	27.3	73.7:26.3
12	79.0	19.3	1.7	92.0: 8.0	21.0	80.5:19.5
13	71.7	27.9	0.4	98.5: 1.5	28.3	72.0:28.0
14	72.1	26.6	1.3	95.5: 4.5	27.9	73.1:26.9
15	70.0	29.1	0.9	97.0: 3.0	30.0	70.7:29.3
Average	75.8	22.4	1.8	91.0: 9.0	24.2	77.2:22.8
16	71.1	28.8	0.1	99.7: 0.3	28.9	72.2:28.8
17	89.0	10.9	0.1	99.0: 1.0	11.0	89.0:11.0
18	0.6	97.0	2.4	97.5: 2.5	99.4	0.8:99.2

* 1–15, perthites; 16, perthite (Hugo pegmatite); 17, microcline (Hugo pegmatite); 18, cleavelandite.

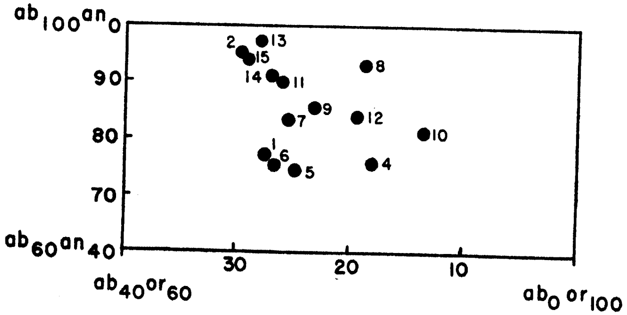


FIG. 2.—The normative ratios Or:Ab and Ab:An of the analyzed feldspars

cryst of the perthitized schist, with the single exception of baria and water, lie within the range of the oxides forming the perthites of the pegmatites. It also shows that the weight percentages of all the oxides of the perthite phenocryst of the granite, with the exception of lime, lie within the range of the oxides forming the perthites of the pegmatites.

TABLE 5

Oxides	Range in the Perthites (Per Cent)	1*	2†
SiO_2	63.89–65.13	65.10	63.96
TiO_2	0.00	0.00	0.00
Al_2O_3	18.92–21.61	19.55	19.39
Fe_2O_3	0.06– 0.13	0.10	0.13
MgO	0.00– 0.12	0.00	0.05
CaO	0.15– 0.82	0.07	0.46
BaO	0.01– 0.06	0.02	0.17
Na_2O	0.38– 3.29	2.98	2.80
K_2O	10.58–15.62	11.58	12.41
H_2O^+	0.12– 0.31	0.16	0.36

* Perthite phenocryst in granite.
† Perthite metacryst in schist.

TEXTURES OF PERTHITES

The analyzed perthites as well as a few others were examined microscropically. Four textural types have been recognized, namely,

1. Patch perthite
2. Vein perthite
3. Film perthite
4. String perthite

These types were named by Andersen (1928), who gave a complete description of each type. The interlocking perthite of Andersen (1928, p. 151), the braid perthite of Goldich and Kinser (1939), and the guttate perthite found by Barth (1930) were not observed in the Black Hills perthites. Two or more types of perthite always occur together. Table 6 shows the types found in the perthites analyzed. Those not analyzed chemically are similar to those analyzed.

Vein and patch types of perthite are abundant in practically all material studied. The patches reach 2.3 mm. or more in width and in most cases exceed 1.0 mm. In the patches of albite are islands of microcline, which exhibit complete optical continuity with the host microcline (pl. 2, *O*). They are well shown in sections parallel to both (010) (pl. 2, *E*) and (001) (pl. 2, *F*). Plates 2, *D–F*, and 3, *A* and *B*, illustrate these relations. In sections parallel to (001), regular bands of albite are commonly found which are parallel to the pericline twins of microcline and in many cases swell out perpendicular to this direction. Commonly the patches send out branches or offshoots which appear to have eaten their way into the microcline. These are irregular and in places have a complicated texture. In some cases the patches form lens-shaped or roughly rounded pools, which send off other patches which may run in any direction. The contacts between the albite of the patches and the microcline are commonly irregular and embayed (pl. 2, *E*). The amount of albite in the patches, estimated by microscopic areal measurements, is not the same in all perthites. In perthite No. 10 albite patches form about 8 per cent, whereas in perthite No. 11, albite patches form approximately 19 per cent of the whole specimen. Patch albite commonly shows polysynthetic twinning according to the albite law. The extinction angle on (010) varies, probably because of variation in the anorthite content. The microcline portions of the perthites are more homogeneous. The extinction angles of microcline are +15° to +16° on (001) and +5° to +6° on (010). The grating may differ in coarseness; in some cases it is fine, and in others it is relatively coarse. A combination of both fine and coarse gratings in a single perthite is common. Nearly all the microcline seen contains some areas, variable in width, which lack the characteristic grating (pl. 2, *D*). Some graphic granites, composed essentially of quartz and perthite, also show a patch type which displays the general characteristics mentioned above.

The vein type of perthite is closely associated with the patch type. In some places it is difficult to differentiate between them. The vein perthite generally shows intricate and irregular networks of anastomosing veins, which grade into the patch type. The veins commonly enclose microcline islands in complete optical continuity with the host microcline similar to those in the patch type. Albite of the vein type rarely exhibits polysynthetic twinning, and its edges are commonly irregular. Both the width and the length of the albite of the vein type vary in the different perthites, but they are decidedly less than those of the patches. The width varies from 0.05 to 0.2 mm., and the length reaches a few millimeters.

TABLE 6

TEXTURAL TYPES OF ANALYZED PERTHITES

No.	String	Film	Vein	Patch	Color*
1	T†	T	X	X	Pale pink
2	X	T	T	X	Grayish-white
3	T	T	X	X	Pink
4	...	...	X	X	Grayish-white
5	T	X	T	X	Deep pink
6	T	T	X	X	Pink
7	T	X	X	T	Salmon-red
8	...	X	X	T	Whitish-gray
9	X	T	X	X	Grayish
10	...	T	X	T	Grayish-white
11	T	T	X	X	Deep pink
12	T	...	X	X	Whitish-gray
13	T	T	X	X	Pink
14	X	...	X	T	Deep pink
15	...	...	X	X	Pinkish-gray

* The perthites analyzed vary in color. The percentage of Fe_2O_3 may be the same in both pink and gray perthites. For example, perthites 1 and 2 are pink and grayish-white, respectively; yet each has 0.09 per cent Fe_2O_3. This indicates that the variation in color is not due to iron or apparently to any of the common elements but might be attributed to other rarer elements.

† Symbols: X = Present in abundance; T = Present in trace; . . . = Absent.

The veins in some cases are more or less parallel to (100) (pl. 3, *C*), but more commonly they run irregularly through the whole perthite. They generally have numerous veinlets which possess smaller width and length. Veins which have circular cross-sections of a diameter of about 0.05 mm. are also present (pl. 3, *D*).

The film and string albites constitute a relatively small amount of the total albite of the perthite, or both films and strings may be absent. The strings and films are always accompanied by vein and patch types (pl. 3, *A*). Both of the latter form, on the average, 90 per cent of the total albite of the perthite, the rest being film and string types. The total albite, estimated by microscopic areal measurements, conforms closely to the normative plagioclase percentages of the various analyzed perthites given in table 4. The films are generally 0.01 mm. in width, rarely 0.04 mm., and several millimeters long. The strings have a width of about 0.003 mm. and a length of as much as 0.65 mm. Both film and string types commonly grade into the vein type. The strings make an angle of 76°–81° with (001). They commonly occur in groups between the veins. The grating of the microcline of the strings is relatively fine, but such fine grating can also be found in the other types.

SIGNIFICANCE OF PERTHITE TEXTURES

The perthitic intergrowth has been thought to be due to (1) simultaneous crystallization, (2) exsolution, or (3) replacement.

A detailed study of the textural features of perthites of southern Norway led Andersen to believe that each well-defined textural class corresponds to a particular mode of origin (1928, p. 163). The string and film types were believed to be the products of exsolution. The string forms at an early stage of the evolution of the feldspar, whereas the films form at a lower temperature in a later stage. The vein perthite was believed to be formed mainly by recrystallization of albitic solutions derived from the pegmatitic magma from which the host feldspar crystallized. These solutions fill the contraction cracks, believed by Andersen to be formed perpendicular to (010) and making an angle of about 70° with (001), where they crystallized without much replacement. Goldich and Kinser (1939, p. 423) thought the atomic structure was an important factor in determining the location of the openings in which the vein solutions would crystallize. In a later stage, Andersen says, these solutions work their way farther from the cracks, replacing the host microcline and forming networks of albitic material. The patch type was considered to be an advanced stage of replacement, transitional between the vein type and pure albite.

It is reasonable to accept the replacement hypothesis for the explanation of the origin of the patch type. It is difficult to believe that magmatic crystallization can give rise to the peculiar characteristics of this type. The microcline islands in complete optical continuity with the host microcline, the extreme irregularity of the contact between the patches and the host, the numerous embayments, the network-like forms and other complex textures without any regularity in the direction of elongation, the transition of veins into patches and of the latter into chessboard albites and albite in general, the variation in the amounts of albite of the patches of different perthites, and the pools and pods of albite with offshoots and tongues which appear to have worked their way by eating and digesting the host—all these features undoubtedly favor the replacement theory.

Vein perthite, which is next to the patch type in abundance, commonly shows microcline relics and irregular and embayed contacts with the host microcline. Intricate networks of arborescent and anastomosing veins are common. These generally grade into the patch type. It is, therefore, a product of metasomatic processes, that is, a crystallization of albitic materials circulating into the contraction cracks and/or the openings controlled by the atomic structure of the feldspar. Much replacement of the potash feldspar accompanies its precipitation.

Andersen's concept of origin of string and film types of perthites seems correct. These perthites are the rarest types. They are believed formed by exsolution in the solid state, that is, by diffusion of the constituents of albite to cracks in the host microcline. Both types show more uniform elongation than do the vein and patch classes. Transitional stages between string and film types, on the one hand, and the vein type, on the other, are not uncommon. Some veins cut the strings and films and are therefore younger than such strings and films. In the early stages of the evolution of the perthite, diffusion of the albitic material seems to have followed a certain preferred direction in the atomic structure of the feldspar forming the regularity in trend of the strings and films. Later in the evolution some of the albitic materials acted selectively in certain directions in the feldspar, giving rise to the regular veins. They might have started this action in the same direction as those of the strings and films, transforming these into the vein type. Perthite No. 15 is devoid of both the string and the film types. This may be because the transformation of strings and films into veins has been completed.

What is the origin of the homogeneous potash-soda feldspar and of the circulating albitic material? It has been proved experimentally that soda feldspar dissolves to a considerable extent in potash feldspar by prolonged heating somewhat below the melting temperatures (Dittler and Köhler, 1925). It has also been proved by X-ray analyses that homogeneous potash-soda feldspar may form cryptoperthite on heating slightly below its melting temperature (Kozu and Endo, 1921; Chao, Smare, and Taylor, 1939). It may be concluded, therefore, that at relatively high magmatic temperatures the potash-soda feldspar is homogeneous. On cooling, the solubility of the soda feldspar decreases rapidly, so that a once homogeneous substance separates into two solid phases, namely, a soda-rich and a potash-rich feldspar. The soda feldspar in this case would be found in fine or thin intergrowths, in other words, strings and films, but not in relatively coarse bands, like those of the veins and patches. The perthite obtained experimentally is a submicroscopic cryptoperthite having a greater similarity to the strings, and to a less extent to the films, than to any of the other classes.

Although the string perthite can be produced by exsolution from magmatic sources, some may have been formed by migration of material from the country rock to structurally favored places where they form rocks markedly unlike the original parent-rock. This process is described elsewhere in this paper. The perthite of the perthitic schists (pl. 1, *A–D*) could not be derived from a magmatic source. It shows megascopic and microscopic features, discussed above, which are difficult to explain by viscous intrusion and magmatic crystallization.

PERTHITE GENESIS

Magmatic pegmatites must, by definition, precipitate from the residual magmatic melt and are its final product of differentiation. The water content of this residual melt was probably less than 10–12 per cent. Such melts should obey the laws of crystal fractionation.

The concentration of lithium or sodium in certain pegmatites and the impoverishment of others in potassium seem to indicate that these rocks might not have been derived from the crystallization of a pegmatitic rest magma. The residual magma is known to be rich in the alkalies in the latest stages, and there is no way of separating the different alkali members by the differential crystallization of the liquor, letting one or another of them concentrate in certain more favorable structures.

In some places, as in the case of the Edison pegmatite in the Keystone area, potassium-bearing minerals are not found. This pegmatite has an interior zone of quartz and albite, surrounded by a zone of quartz, albite, and spodumene. In other cases, such as the Etta pegmatite of the Keystone area, in which spodumene, the dominant mineral, forms crystals several feet in length, lithium is present in abundance; in still other pegmatites, lithium is found only in traces.

Whenever perthite forms pegmatitic pockets or forms a zone in a pegmatite body, its chemical composition should closely approximate the mother-liquor from which it was crystallized. Walker and Mathias (1946) believe that the perthite at Bantry Bay (South Africa) represents the composition of the pegmatitic magma. If perthites crystallize from a magma, such magma should follow the course of crystallization expressed by the feldspar equilibrium diagram.

The orthoclase-albite equilibrium diagram suggested by Vogt (1926) does not satisfactorily explain the derivation of the perthites analyzed from a granitic magma. The orthoclase:albite ratio in these perthites (table 4) is far from that which should be attained (that is, Or 40:Ab 60) had they been formed in such a manner.

Spencer (1938) in his study also concluded that the orthoclase-albite binary diagram cannot explain satisfactorily the magmatic origin of the potash-rich pegmatites.

THE ORTHOCLASE-ALBITE-ANORTHITE TERNARY EQUILIBRIUM DIAGRAM

The ratio Or:Ab:An of the different perthites given in table 4 was plotted on the orthoclase-albite-anorthite equilibrium diagram as shown in figure 3. The cotectic curve, CD_1, in figure 3, after Bowen (1928), has C at the Or:An eutectic point and D at the Ab:Or eutectic point.

All the perthites analyzed lie in the field of the orthoclase or the potash-rich rocks. Those of Andersen (1928, p. 148) fall in the same field. In the normal course of crystallization, composition of the liquid cannot cross the cotectic curve. In the plagioclase field the crystallizing liquid must follow a certain path, such as that given by Nockolds (1946, p. 215) and plotted in figure 6 or that given by Barth (1944). The liquid, upon hitting the cotectic curve, would then follow it, so that in the last stages of crystallization the eutectic ratio Or:Ab would be attained. The Or:Ab:An ratio of the Black Hills rocks does not harmonize with this concept.

The author also plotted the Or:Ab:An content of some fifty rocks.[4] These are deep seated (graphic granites, granites, pegmatites, and a monzonite), hypabyssal (porphyries), and extrusive (rhyolites, liparites, pitchstones, comendites, and trachytes). The compositions of these fifty rocks also lie in the orthoclase field (fig. 4). An attempt to shift the Or:Ab eutectic point D of figure 4 to D^1, around which most of the plotted graphic granites and granites fall, and the Or:An eutectic point C to C^1, which represents the composition of a pitchstone,[5] would not solve the problem, as there are still other rocks[6] whose composition still lies in the orthoclase field (fig. 5). Why should these rocks, if of magmatic origin, not follow the regular course of crystallization that such an origin demands according to the feldspar ternary diagram?

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[4] Of subrang Dopotassic Omeose, Order Quardofelic Britannare, and the subrang Dopotassic Magdeburgose and Sodipotassic Alaskose, Order Quarfelic Columbare, from Washington's tables (1917, pp. 109, 57, and 61, respectively).

[5] Belonging to the subrang Perpotassic, Order Quarfelic Columbare (Washington, 1917, p. 79).

[6] Of the subrang Dopotassic Omeose and the subrang Perpotassic Lebachose, Order Quardofelic Britannare (Washington, 1917, p. 107).

Bowen has proved (1937) that liquids at the latest stages of crystallization become enriched in the alkali silicates. The chemical analyses of some volcanic rocks have been shown to fall in the trough of his figure 9 (1937, p. 18), which he postulates for rocks formed in the last stages of crystallization. Some deep-seated rocks (granites and syenites) have been shown to lie also in this trough. However, most of the potash-rich rocks—extrusive as well as deep seated—do not lie in this trough. With respect to the origin of such rocks, Bowen referred to the work of R. Terzaghi (1935). She stated that some of the potash-rich rocks may have been formed by metasomatic alterations but that most of them are magmatic. The position of the latter rocks in the orthoclase field Or:Ab:An ternary diagram she attributes mainly to the reaction relation between anorthite and orthoclase, to pressure, or to devitrification.

The reaction relation between anorthite and orthoclase as suggested by Bowen (1928, pp. 227–233) was thought

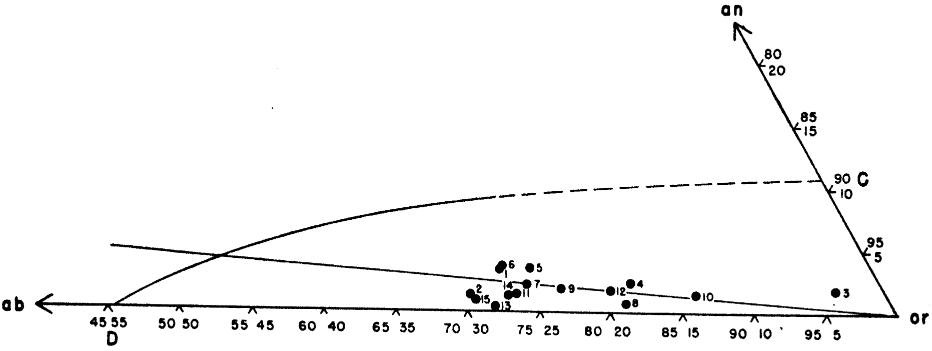

FIG. 3.—The Or:Ab:An ratio of the analyzed perthites

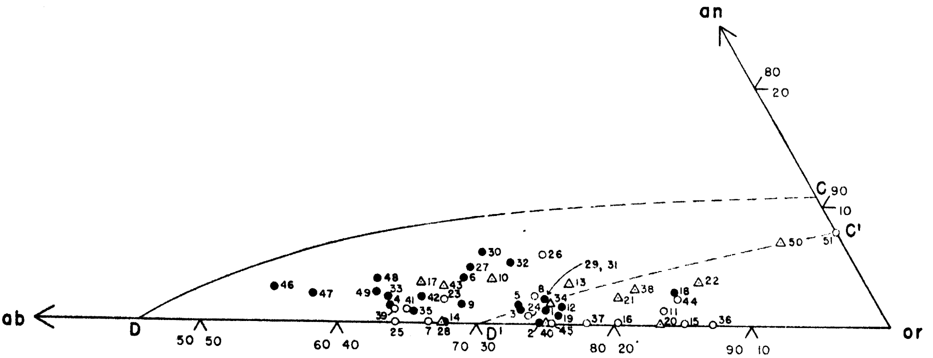

FIG. 4.—The Or:Ab:An ratio of some potash-rich rocks. Nos. 1–5 and 6–32 = nos. 1–5 and 7–33 of Washington (1917, p. 109). Nos. 33–45 = nos. 1, 2, 3, 11, 12, 13, 15, 18, 22, 25, 27, 30, and 31 of Washington (1917, p. 57). Nos. 46–49 = nos. 10, 31, 32, and 60 of Washington (1917, p. 61). Nos. 50–51 = nos. 1–2 of Washington (1917, p. 79). Solid circle = granite, pegmatite; open circle, rhyolite, trachyte, and other volcanics; triangle, porphyry.

to enable the liquids to cross the cotectic curve. Recently, Schairer and Bowen (1947), in their study of the system anorthite-leucite-silica, found that such a reaction relation does not exist. There is, however, a eutectic between anorthite and orthoclase at 1040° ± 20° C. Therefore, the reaction factor can no longer be held responsible for liquids crossing the cotectic curve.

Terzaghi (1935) says that an increase of pressure under deep-seated conditions can shift the orthoclase-albite eutectic point; she admits, however, that, in order to establish this, more information is needed concerning the latent heat of formation of the feldspars. If such were the case, the plagioclase field would increase, and some of the potash-rich rocks would no longer lie in the orthoclase field. But, as has been mentioned above (p. 566), such a shift, on the basis of the position of most of the graphic granites (which closely approximates the amount of shift postulated by Terzaghi), is not adequate because other potash-rich rocks will still remain in the orthoclase field. The effect of pressure on the feldspar equilibrium diagram appears to be of little importance because some volcanic rocks, such as rhyolites and liparites, as well as some deep-seated granites and pegmatites, lie together in the same field. Furthermore, Nockolds (1946) found that there are deep-seated aplites, adamellites, and granites which lie on the cotectic curve and follow its course (fig. 6). The Or:Ab:An ratios of some fifty deep-seated, hypabyssal, and extrusive rocks,[7] plotted in figure 6, lie approximately on the cotectic curve, and some of them fall in close proximity to the Or:Ab eutectic point. Such types of rocks might be mainly of magmatic origin. Furthermore, there should be pegmatites corresponding to such rocks with compositions approximately on the cotectic curve, so that the position of this curve as given by Bowen in the orthoclase-albite-anorthite equilibrium diagram seems to be approximately correct.

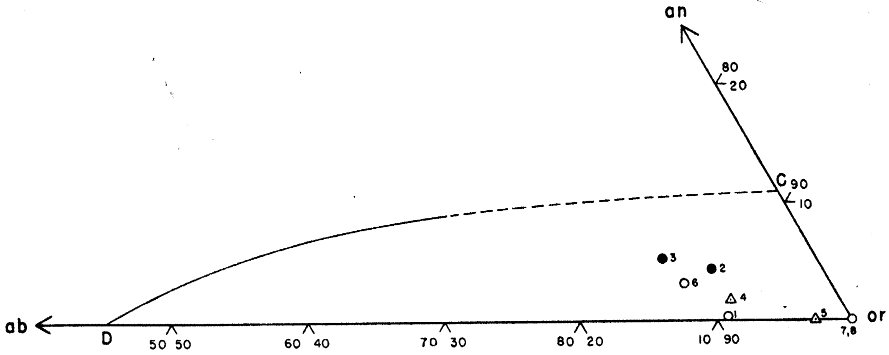

FIG. 5.—The Or:Ab:An ratio of some potash-rich rocks. Nos. 1–8 = nos. 4, 5, 7, 8, 9, 10, 11, and 12 of Washington (1917, p. 107).

Therefore, if pressure was not the factor responsible for throwing potash-rich rocks into the orthoclase field, would it not also have affected Nockolds' selected rocks and the others of figure 6 which do not lie in that field? Why, then,

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[7] Of the subrangs Sodipotassic Liparose, Dopotassic Mihalose, and Dopotassic Dellenose (Washington, 1917, pp. 145, 79, and 165, respectively).

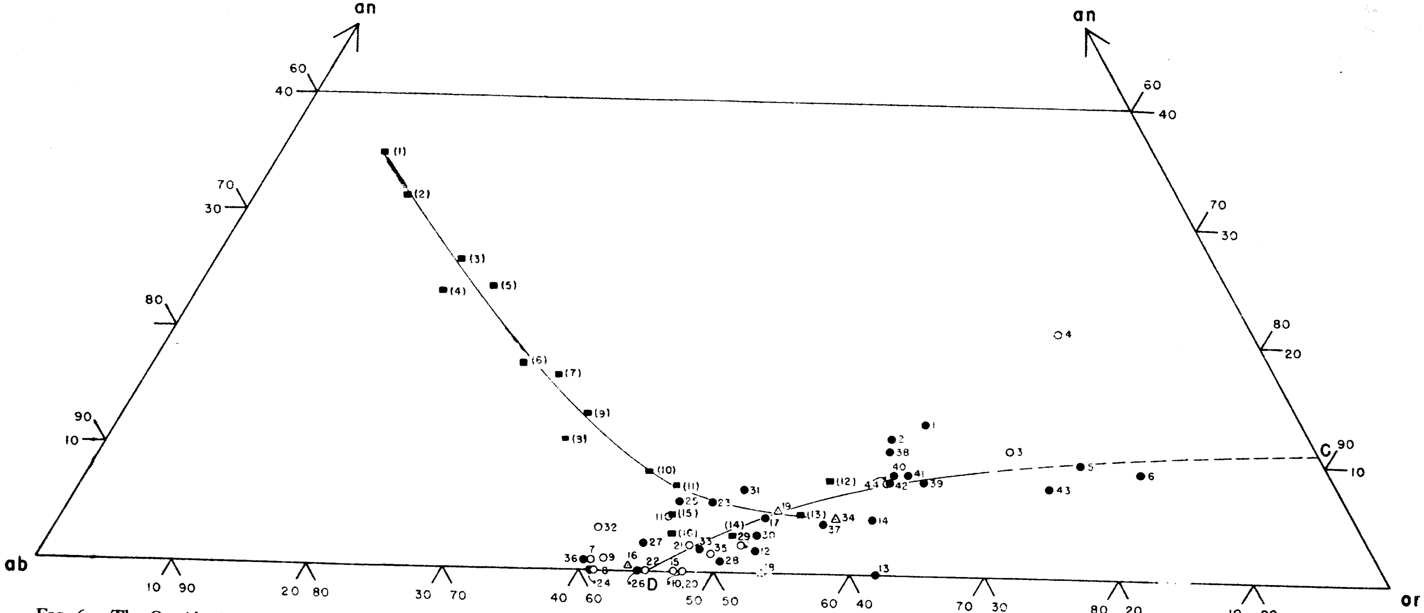

FIG. 6.—The Or : Ab : An ratio of some rocks following a course of crystallization in the plagioclase field and of rocks falling approximately on the cotectic curve. Nos. (1)–(16) = nos. 1–16 of Nockolds (1946). Nos. 1–6 = nos. 3–8 of Washington (1917, p. 79). Nos. 7–37 = nos. 248, 249, 250, 254, 255, 256, 257, 258, 260, 261, 262, 263, 264, 265, 267, 270, 276, 279, 280, 282, 283, 284, 285, 289, 290, 291, 292, 294, 296, and 297 of Washington (1917, p. 145). Nos. 38–44 = nos. 1, 5, 7, 8, 10, 17, and 22 of Washington (1917, p. 165). Square, granite, adamellite, and aplite of Nockolds (1946).

are they still governed by the magmatic rules? Why should pressure have affected some rocks and not others? If one argues that rocks studied by Nockolds, though deep seated, were subjected to less pressure than were the deep-seated rocks in the orthoclase field, then it follows that the volcanic rocks in the orthoclase field also were formed under conditions of higher pressure than the rocks of Nockolds—a conclusion obviously absurd. There probably are deep-seated rocks—granites and aplites, as well as pegmatites—which are unquestionably of magmatic origin; but these will generally have normative feldspar contents which follow the cotectic curve, as in the case of those rocks selected by Nockolds. At the same time, there are other deep-seated varieties of the same species which do not follow the magmatic crystallization rule. The orthoclase-albite-anorthite ratios of such rocks cannot result from strictly magmatic processes; thus metasomatic alterations must have played the significant role in their formation.

It is important to note that the zoned plagioclase crystals of an extrusive rock formed under low pressure, such as a basalt, and those of a deep-seated rock formed under relatively high pressure, such as a gabbro, show agreement with the course of crystallization given in the the albite-anorthite solid-solution diagram (Bowen, 1928, p. 34). In both rocks the zoned crystals have cores which are richer in the anorthite molecule than are the rims. Moreover, the path of crystallization of both basalts and gabbros terminates at the same point in the diopside-albite-anorthite equilibrium diagram (Bowen, 1928, p. 46). Diopside and albite exhibit quite different crystal structures: chain lattice and framework lattice, respectively. It seems probable that pressure would have a greater effect on systems whose members have different crystal structures. Because no pressure effect is evident in the albite-anorthite and the diopside-plagioclase systems, it must be negligible in the orthoclase-albite-anorthite system.

Terzaghi (1935) found that four devitrified rhyolites were richer in potash and poorer in both soda and lime than were rhyolites not devitrified. The Or: Ab: An ratios of the devitrified rocks fall in the potash-rich or orthoclase field. Terzaghi believed that the devitrified rhyolites were originally magmatic and vitric and that at an earlier stage of their history they had the normal Or: Ab: An ratios of vitreous rhyolites. Later, soda, lime, and potash were leached out—the soda and lime at a greater rate than the potash. Therefore, when these altered rocks became devitrified, they possessed higher potash content than did the original vitric type, and they consequently fell into the orthoclase field. Recently the plot of the normative feldspar content of some rhyolite rocks of the Esterel region in France (Terzaghi, 1948, fig. 1) shows that the vitreous obsidian lies in the plagioclase field, whereas six devitrified obsidians and spherulitic rhyolites and two rhyolite porphyries lie in the orthoclase field. Because devitrification is confined to the immediate vicinity of the cracks in one of the obsidians, Terzaghi believes the devitrification to be due to percolating solutions. The crystals of alkalic feldspar, which are most strongly euhedral, are believed to have originated by *replacement* of vitreous material. Terzaghi concludes that hot solutions and possibly gaseous emanations were active in the region after the eruptions of rhyolitic lava took place. Reactions between solutions and rocks are most clearly indicated by the microscopic features of the spherulitic rhyolites that she studied. The secondary potash which enriched the rhyolites during their devitrification either was introduced by hydrothermal solutions or gaseous emanations originating from an underlying body of magma or was derived by local leaching of the rhyolite itself. The potash leached from the rhyolite in one place was carried on in solution and was available for potash enrichment elsewhere. The possible dual origin of the secondary potash in the Esterel region is in harmony with the conclusions reached by Allen (1935), Allen and Day (1934), and Fenner (1936) in their studies of the thermal waters and the altered rhyolites of the Yellowstone Park region.

Fenner (1936, p. 278) observed that not only were orthoclase, analcite, heulandite, and calcite formed as replacements of original minerals by soluble salts (chlorides, carbonates, etc., of the alkalies), but all these difficultly soluble minerals were taken into solution to some extent, migrated rather freely, and were deposited in fissures and other open spaces. Elsewhere, Fenner (1936, p. 279) states that there may be many occurrences of similarly altered rocks in volcanic regions, which, in the absence of complete data, have been regarded as normal lavas. Even analyses of such rocks have been accepted as representing original compositions.

The high potash content of some rhyolites and extrusives may be due to secondary enrichment by potash-bearing solutions. The secondary potash might have been derived in part from these extrusives by leaching and migrating from one place to another. Possibly the potash transfer could take place in the solid state, especially so if evidence for the presence of solutions is lacking. If pressure causes differences in the chemical potentials of the elements in the various parts of country rocks, it is believed that the elements will migrate from areas of higher to areas of lower potentials until equilibrium is reattained (Ramberg, 1944).

Other factors may be responsible for potash-rich pegmatites—perhaps the concentration of volatiles in the last stages of crystallization. Andersen (1931, p. 25) gives two analyses of pegmatites whose projection points apparently lie in the plagioclase field. Others can be found in Washington's tables. On the other hand, some pegmatites lie in the orthoclase field. Though all magmatic pegmatites are supposed to have crystallized from residual solutions rich in volatiles, we find some on the cotectic curve, others in the plagioclase field, and still others in the orthoclase field. We can conclude, therefore, only that we have pegmatites of different origins and that in the case of the magmatic types the presence of volatile constituents has had no significant influence on the course of crystallization.

Moreover, the effect of volatile constituents on the course of crystallization is not too well known; their only important effect, however, seems to be a lowering of the temperature of crystallization, with no discernible influence on its course. Bowen states (1928, pp. 299–300) that the addition of 1 per cent H_2O (the most important volatile constituent) to a liquid composed of 50 per cent $K_2Si_2O_5$ and 50 per cent K_2SiO_3 does not change the course of crystallization or even the proportions of the separated phases from that of the dry melt. Moreover, in Bowen's nephelite-kaliophyllite-silica equilibrium diagram (1937), the lowest temperature of crystallization for a dry melt is 1038° C.; but granites rich in the volatiles and thought to have crystallized from a magma at 700° C. have chemical compositions which place them in the "trough" of this dry-melt diagram. The zoned plagioclase crystals of both peridotites poor in volatiles and granites relatively rich in these follow equally well the rules of plagioclase crystallization given by the albite-anorthite binary equilibrium diagram. One is therefore led to the conclusion that, although the presence of the volatile constituents will lower crystallization temperatures, it will not significantly change the course of crystallization. That is, although fusion surfaces might show differences in relief with increase or decrease of volatiles, the positions and relative reliefs of the cotectic "valleys" and boundary surfaces will remain approximately the same.

The immiscibility of water-rich siliceous liquids does not seem to be the solution. The initial compositions used by Tuttle and Friedman (1948) for the determination of the equilibrium relations of the system Na_2O-SiO_2-H_2O at 250, 300, and 350°, proving that the system possesses liquid immiscibility between 200 and 250°, consist of approximately 50 per cent H_2O. This is much higher than the 10 per cent of H_2O assumed to be in the pegmatitic rest magma. Smith (1948, p. 538) also dealt with solutions of a very highly specialized composition that do not represent a rest pegmatitic magma from which a perthitic rock could crystallize. Smith's solutions had 50 per cent H_2O and less than 1 per cent of K_2O. Goranson's work (1931) makes it unlikely that a perthitic liquid which approximates that of a granite could absorb more than 10 per cent H_2O.

Moreover, the potash-rich rhyolites cannot be explained because their temperature of crystallization is higher than that at which immiscibility occurs.

The experiments by Goranson (1931) have demonstrated a limited miscibility of H_2O in silicate melts. A consequence of this is that, near the end of the consolidation history of a magma, a hydrothermal solution may separate as an independent phase coexisting with the magmatic phase, on the one hand, and with the solid phase (the rock), on the other. The significance of Goranson's work and its implications have been discussed by many petrologists. The exudation of the hydrothermal solutions is believed to mark the final stages in the consolidation of the magma. It is important to note that they are separated from the magma through an immiscibility gap. The composition of the solutions is governed by the endomagmatic hydrothermal differentiation processes as described by Neumann (1948). Magmatic pegmatites are the precipitates from the residual magmatic melt prior to the hydrothermal stage.

It appears, therefore, that neither the reaction relation anorthite-orthoclase nor pressure nor the volatile constituents can account for the composition of the Black Hills perthitic pegmatites. Metasomatic alterations which were found by Terzaghi (1948) to be responsible for the presence in the orthoclase field of the potash-rich rhyolites that she studied might also be the reason why the Black Hills potash-rich pegmatites lie in the same field. It is uncertain whether all the potash-rich deep-seated, hypabyssal, and extrusive types plotted (figs. 4 and 5) are metasomatic rocks or were formed by some other processes. This is not discussed in this paper and will be the subject matter of another work by the author.

If it turns out that the hypothesis of potassium metasomatism is true, the orthoclase field of the orthoclase-albite-anorthite ternary diagram might well be named the "potassium metasomatism field." Perhaps also there are rocks which possess a higher albite normative composition than is consistent with a truly magmatic origin. Examples of these are the albite-rich granites and the albitites. A chemical analysis of an albitite from the Black Hills is given in table 3; this has an Or:Ab ratio of 3:97. These would be placed in a field nearer to the Ab corner of the Or-Ab-An ternary diagram; this field might be named the "sodium metasomatism field."

Whether the potash pegmatites of the Black Hills were once magmatic and have been metasomatized or are merely metasomatic is another subject and will be considered elsewhere.

GENESIS OF PERTHITE SCHISTS

Table 3 shows the chemical analyses of three country rocks. The rock formula introduced by Barth (1948) of each of the analyzed rocks—the average perthite, the cleavelandite, and the microcline—was calculated as shown in table 7. Table 8 shows quantitatively the gains and losses required to change the normal schist to a perthitized schist, albitite, and perthite. The normative compositions, using Von Wolff's method, were also calculated in table 9.

Figure 7 is a reproduction of figures 7 and 8 of Reynolds' work on the geochemical changes leading to granitization (1946). The schist of the Glendale area lies in the field of the aureole sediments of the Flamouth granite area. It lies also in the field of the aureole shales of the Dartmoor area (Reynolds, 1946, fig. 5). The perthitized schist falls in the fields representing the Flamouth granite, the granitization series of the Dartmoor area, and the granites of the latter region. The albitite and the average perthite lie in a separate field closer to the *L* of the diagram. This means that partial feldspathization of the schist could give rise to perthitic schist, which has a chemical composition similar to potash granites. If feldspathization is more complete, rocks will form that possess a chemical composition near that represented by *L* of the Von Wolff diagram.

Such rocks are the albitites and the perthitites (monomineralic rocks composed of perthite). The formation of such rocks is an end-stage in Reynolds' sequence of changes of pelitic and semipelitic rocks. These changes may take place by the reorganization of the atoms and molecules inside the schist. Table 8 would then refer to loss and gain inside certain volumes of the schist, whereas the whole schist complex may show little change in bulk chemical composition. The perthite crystals in this schist seem to grow and recrystallize at the expense of the schist material, similar to the formation of porphyroblasts in other metamorphic rocks. In places they seem to have pushed aside the schist which forms the groundmass of the rock (pl. 1, *A*). In places there are crystals exceeding 6 cm. in length, exhibiting megascopic sieve texture (pl. 1, *C*). A perthitic pegmatite which has approximately the same chemical composition and shows identical textures as those of the perthite of the schist may thus be formed in a similar manner. It is worth mentioning that Goldschmidt (1934) attributed the source of scandium, the rare element found in some pegmatitic pockets, to the country rocks which are mainly amphibolites and not to a pegmatitic magma. This was concluded from spectrographic studies. The amphibolites which could give their scandium lend support to the belief that the schist could contribute potassium to form the feldspars. If this be true, the change is illustrated by the following: Three schist "unit cells" (the least number of the schist "unit cell" to form a perthite is 3) are composed of: K 16.86, Na 5.37, Ca 1.62, Mg 8.76, Fe 11.43, Al 45.45, Ti 0.90, Si 181.35, P 0.39, H 26.85; and the average perthite "unit cell" consists of: K 14.45, Na 4.35, Ca 0.39, Mg. 0.05, Fe 0.07, Al 21.20, Ti 0.0, Si 58.80, H 1.37.[8] Therefore, the equation of the process of transformation will be

$$3 \text{ schist} \rightleftharpoons \text{perthite} + 2 \text{ schist relic.}^{9} \quad (1)$$

The "unit cell" of this schist relic is composed of: K 1.21, Na 0.51, Ca 0.62, Mg 4.36, Fe 5.68, Al 12.13, Ti 0.45, Si 61.28, P 0.20, H 12.74 (see Table 10).

TABLE 7

FORMULAE OF THE COUNTRY ROCKS

Rock	K	Na	Ca	Mg	Ba	Mn	Fe^{++}	Fe^{+++}	Al	Ti	Si	P	O	H
Schist	5.62	1.79	0.50	2.92	0.04	0.04	2.82	0.95	15.15	0.30	60.45	0.13	160	8.95
Perthitic schist	7.20	3.80	0.38	1.46	0.04	0.00	0.78	0.42	15.70	0.16	61.62	0.32	160	6.85
Albitite	0.48	16.30	0.54	0.70	0.00	0.00	0.15	0.10	18.34	0.00	60.51	0.38	160	1.32
Average perthite	14.45	4.35	0.37	0.05	0.02	....	....	0.07	21.20	0.00	58.80	....	160	1.37
Cleavelandite	0.11	18.37	0.47	0.11	0.01	....	....	0.06	20.24	0.00	59.21	....	160	2.42
Microcline	17.50	2.15	0.02	0.03	....	....	....	0.03	20.90	0.00	59.00	....	160	1.65

TABLE 8

GAINS AND LOSSES IN THE SCHIST

Rock	Addition and Subtraction
Schist to perthitic schist	*Addition:* 1.58 K, 2.01 Na, 0.55 Al, 1.17 Si, 0.19 P *Subtraction:* 0.12 Ca, 1.46 Mg, 0.04 Mn, 2.57 Fe, 0.14 Ti, 2.10 H
Schist to albitite	*Addition:* 14.51 Na, 0.04 Ca, 3.19 Al, 0.06 Si, 0.25 P *Subtraction:* 5.14 K, 2.22 Mg, 0.04 Ba, 0.04 Mn, 3.52 Fe, 0.30 Ti, 7.63 H
Schist to average perthite	*Addition:* 8.83 K, 2.56 Na, 6.05 Al *Subtraction:* 0.13 Ca, 2.87 Mg, 0.02 Ba, 0.04 Mn, 3.70 Fe, 0.30 Ti, 1.65 Si, 0.13 P, 7.58 H

TABLE 9

NORMATIVE COMPOSITION AFTER VON WOLFF

Rock	Alk *f*	An	Mt	Aug	*Q*	*t*	*L*	*M*	*Q*
Schist	37.6	2.4	1.2	13.4	41.1	4.3	44.3	14.6	41.1
Perthitic schist	55.2	2.0	0.6	5.2	34.3	2.8	60.0	5.8	34.3
Albitite	84.2	2.8	0.1	2.0	10.9	0.2	87.2	2.1	10.9
Average perthite	94.7	3.9	0.1	0.1	0.1	1.1	99.7	0.2	0.1

Figure 8 shows a crystal of perthite about 2 cm. in length, forming a zone surrounded by another zone containing schist relic material. Figure 9 is drawn to explain a mode of formation of such a crystal. The arrows indicate the direction of migration, and the amounts of the diffusing ions are shown. The three "unit cells" of equation (1) are called in figure 9 the "outer shell," the "inner shell," and the "core." Potassium and sodium migrated from both the outer and the inner shells toward the core, which acted as a receiver for the feldspathic material. The core, which possesses K 5.62 and Na 1.79, has thus gained K 5.62 and Na 1.79 from the inner shell and K 3.21 and Na 0.77 from the outer shell. The core also gained Al 6.05 migrating from the inner shell and at the same time lost the excess of other ions over and above those necessary to form the perthite. The inner shell has gained K 1.21, Na 0.51, and Al 3.03 migrating from the outer shell and the other ions

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[8] Ba is taken into Ca, and Mn is taken into Fe.

[9] Material produced after the petroblastic formation of perthite from the schist.

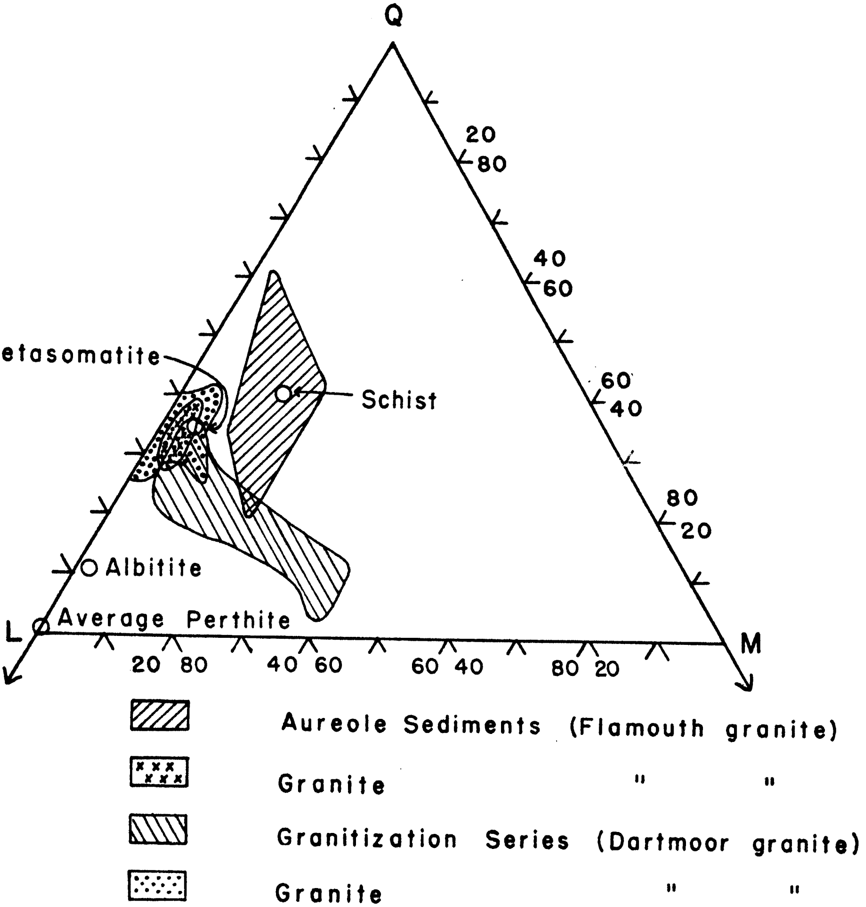

FIG. 7.—Von Wolff diagram showing the composition of the studied rocks

migrating from the core in the amounts indicated on the arrows. Lastly, the outer shell gained material migrating from the core as given in figure 9. In this case it is supposed that the outer and the inner shells have the same composition after the development of the perthite. If such were not the case, the outer and inner shells would form two zones different in composition. The development of a pegmatite body composed almost entirely of perthite surrounded by schist material may be the result of such a transformation process.

The perthite zone of a pegmatite might form in a similar manner. But equation (1) does not apply if several zones are present. If the ratio and chemical composition of the different minerals forming these zones are known, one can set up a similar transformation equation.

The composition of cleavelandite (table 2) is considered to represent that of albite. It is possible for both microcline and albite to form from the schist according to the following equation (illustrated in table 11):

$$16 \text{ schist} \rightleftharpoons 4 \text{ microcline}^{10} + \text{albite (cleavelandite)} + 11 \text{ schist relic} . \qquad (2)$$

Sixteen schist "unit cells" are the least number of these units which could form a potash-rich and a soda-rich feldspar in the approximate ratio found in the average perthite, that is, 4:1, respectively.

In this case, if conditions in the country rock were favorable for zoning, the schist could give rise to a pegmatite with two zones, one composed of microcline and the other of albite. This transformation might take place at lower temperatures than those of the formation of perthite. This is because, at relatively high temperatures, the feldspathic material is likely to form a homogeneous potash-soda feldspar. It should be noted that the formula of the schist relic of the last equation is approximately the same as that of the schist relic derived from equation (1).

TABLE 10

	3 Schist =	Perthite +	2 Schist Relic	Formula of Schist Relic
K	16.86	14.45	2.41	1.21
Na	5.37	4.35	1.02	0.51
Ca	1.50	0.39	1.13	0.57
Mg	8.76	0.05	8.71	4.36
Ba	0.12	0.02	0.10	0.05
Mn	0.12	....	0.12	0.06
Fe^{++}	8.46	....	8.46	4.23
Fe^{+++}	2.85	0.07	2.78	1.39
Al	45.45	21.20	24.25	12.13
Ti	0.90	0.00	0.90	0.45
Si	181.35	58.80	122.55	61.28
P	0.39	....	0.39	0.20
H	26.85	1.37	25.48	12.74

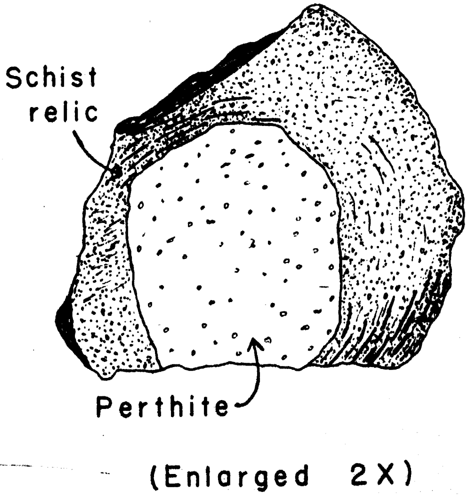

FIG. 8.—Perthite metacryst in the schist

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[10] The formula of microcline (table 1) of Hugo pegmatite is used here.

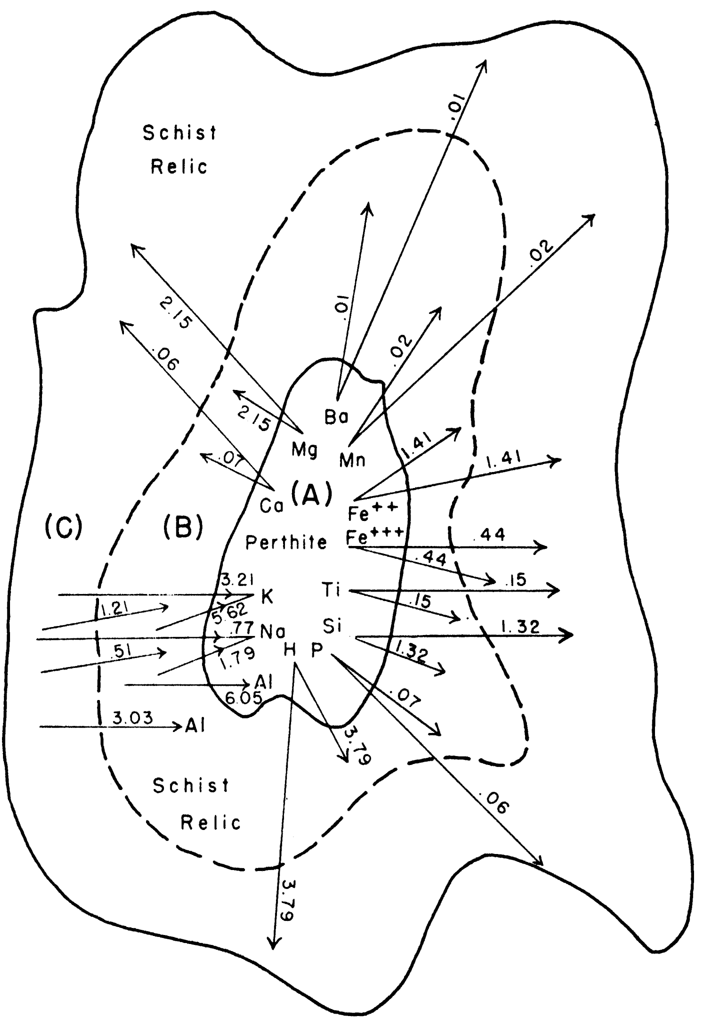

FIG. 9.—Reorganization of the schist material to form the perthite metacryst. *A*, core of innermost shell; *B*, inner shell; *C*, outer shell.

It is worth mentioning that the formation of microcline and albite suggests an explanation of the plagioclase rims surrounding orthoclase in some potash granites and the orthoclase rims surrounding plagioclase in some basic rocks. Such rimmed plagioclases were considered by Bowen (1928, pp. 227–233) as evidence for the existence of a reaction relation between anorthite and orthoclase. In a recent work, Schairer and Bowen (1947) found that anorthite and orthoclase are not reaction pairs; therefore, Bowen's first conclusion is no longer valid, and the rimmed plagioclases of the basic rocks must now be explained in another way. They could find their solution in solid diffusion phenomena.

TABLE 11

	16 Schist	= 4 Microcline	+ Cleavelandite	+ 11 Schist Relic	Formula of Schist Relic
K	89.92	70.00	0.11	19.81	1.80
Na	28.64	8.60	18.37	1.67	0.15
Ca	8.00	0.04	0.47	7.49	0.67
Mg	46.72	0.12	0.11	46.49	4.22
Ba	0.64	....	....	0.64	0.06
Mn	0.64	....	....	0.64	0.06
Fe^{++}	45.12	....	....	45.12	4.10
Fe^{+++}	15.20	0.12	0.06	15.02	1.40
Al	242.40	83.60	20.24	138.60	12.60
Ti	4.80	0.00	0.00	4.80	0.43
Si	967.20	236.00	59.21	671.99	61.09
P	2.08	....	....	2.08	0.19
H	143.20	6.60	2.42	134.20	12.18

If microcline (potash-rich material) and albite (soda-rich material) derived from the schist subsequently reacted metasomatically by replacing the schist, other rocks would result, according to the following equation (illustrated in table 12):

$$\text{Microcline} + \text{albite (cleavelandite)} + \text{schist} \rightleftharpoons \text{perthite} + \text{albitite} + \text{perthite schist} + \text{basic ions}\,. \tag{3}$$

TABLE 12

	Microcline	+ Cleavelandite	+ Schist	= Perthite	+ Albitite	+ Perthite Schist	+ Basic Ions
K	17.50	0.11	5.62	14.45	0.48	7.20	....
Na	2.15	18.73	1.79	4.35	16.30	3.80	....
Ca	0.02	0.47	0.50	0.37	0.54	0.38	....
Mg	0.03	0.11	2.92	0.05	0.70	1.46	0.85
Ba	....	0.01	0.04	0.02	0.00	0.04	....
Mn	....	....	0.04	....	0.00	0.00	....
Fe^{++}	....	....	2.82	....	0.15	0.78	1.89
Fe^{+++}	0.03	0.06	0.95	0.07	0.10	0.42	0.45
Al	20.90	20.24	15.15	21.20	18.34	15.70	....
Ti	0.00	0.00	0.30	0.00	0.00	0.16	0.14
Si	59.00	59.21	60.45	58.80	60.51	61.62	....
H	1.65	2.42	8.95	1.37	1.32	6.85	....

The amounts of potassium, sodium, aluminum, and silica of the reactants are approximately equal to those of the products; however, basic ions are released in the reaction and are thus free to migrate to other parts of the country rock, where they may, by reaction, give rise to other basic rocks. Perhaps they migrated to the north of the studied area, where they may have played a role in the formation of some amphibolites outcropping there. Another working hypothesis is that they might have diffused to lower levels in the gravitational field. A combination of both explanations is also a possibility. Figure 10 is a summary of the reactions discussed above.

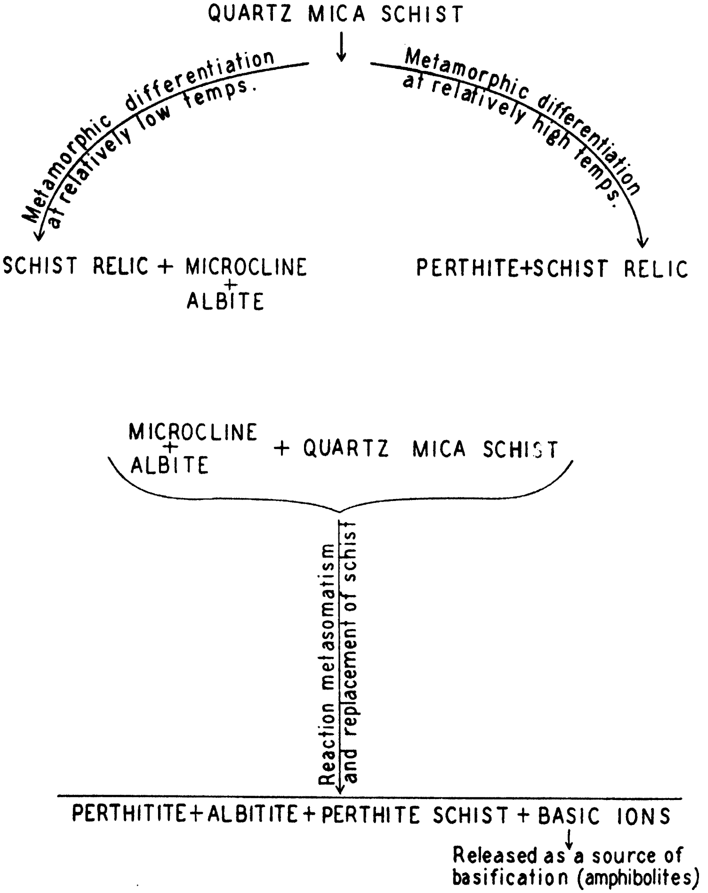
QUARTZ MICA SCHIST
Metamorphic differentiation at relatively low temps.
Metamorphic differentiation at relatively high temps.
SCHIST RELIC + MICROCLINE + ALBITE
PERTHITE+SCHIST RELIC
MICROCLINE + ALBITE + QUARTZ MICA SCHIST
Reaction metasomatism and replacement of schist
PERTHITITE+ALBITITE+PERTHITE SCHIST + BASIC IONS
Released as a source of basification (amphibolites)

FIG. 10.—Metasomatic development of the Black Hills perthite

SUMMARY AND CONCLUSIONS

Some pegmatitic perthites and country rocks of the Glendale area of the Black Hills of South Dakota were chemically analyzed and microscopically studied in order to explain the origin of the perthites of this area. The textures of the perthites and the microscopic features of the country rocks do not seem to be of a magmatic nature; rather, they support metasomatic activities. Evidence of metasomatic derivation of the perthites was given in the discussion of their chemistry. The feldspar equilibrium diagram was proved to be unsatisfactory for the magmatic derivation of these perthites. Moreover, the general chemistry does not harmonize with the general crystallization rules. The schists grade into feldspar-rich types, the country schists having been chemically reactivated. The feldspathic materials are thought to be endogenetically derived from the schist. Perthitization was partial in some parts of the country rock, where perthite schists were developed. In other parts of the country rock perthitization was complete because of favorable conditions of temperature, pressure, and structure, and thus perthitic pegmatite bodies formed. In still other parts of the country rock the conditions assisted both in the trapping of some elements and in the action of certain reaction processes which gave rise to the formation of shells or zones composed essentially of definite minerals characteristic of such conditions. The perthite zone of certain pegmatities is considered to be one of these zones.

The formation of perthite in general was a complex metasomatic process which took place in various stages. The following steps are believed to have taken place during the evolution of the studied perthites: (1) endogenetic development of homogeneous potash-soda feldspathic material by metamorphic differentiation of the country rocks; (2) diffusion of the soda feldspar portion into preferred parts of the structure of the homogeneous material to form the string and film types; (3) endogenetic development of a soda-feldspar (albite) from the schist; and (4) reaction metasomatism or replacement of microcline of the film and string types of perthite, giving rise to vein and patch classes. This reaction was accelerated by water derived from the schist.

I do not pretend to have discussed the whole problem of the genesis of pegmatites. I have given only some explanations of the studied types.

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SANDSTONE TYPES: THEIR ABUNDANCE AND CEMENTING AGENTS[1]

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ABSTRACT

A selected sample of 277 sandstones (183 from thin sections and 94 from descriptions in the literature) were classified into five general classes on the basis of detrital mineral composition. The relative abundance of each of these classes was determined, as well as the average mineral composition of each class. A comparison of the carbonate-cemented sandstones with the silica-cemented varieties showed an increase in carbonate in later geologic periods. In the carbonate-cemented sandstones, partial replacement of the detrital grains was almost universal.

INTRODUCTION

This study had two objectives. First, an attempt was made to measure the relative abundance of each of the major types of sandstones. Second, the variation with respect to time of the ratio of the silica-cemented sandstones to those carbonate-cemented, was studied.

Thin sections of the specimens were examined microscopically to determine specifically the detrital mineral components and the cementing agent and, incidentally, to record some of the structure of the rock, including such phenomena as replacement.

ACCUMULATION OF THE DATA AND METHOD OF TABULATION

The initial problem of this study was to obtain a sample which would be representative of all sandstones. In an effort to accomplish this, the author secured 277 specimens of wide geographic and stratigraphic range from the United States, Canada, Alaska, and Europe. Nevertheless, certain areas were better represented than others. This was due largely to the greater geologic exploration and activity that has taken place in these areas. The extent to which this invalidates the sample is unknown, but it is believed that a sufficient diversification has been obtained to make these effects minor. An attempt was made to use only such sandstones as might be considered typical of their particular locality of origin.

The data for this study were obtained from four sources. Three of these sources were collections of rock specimens which were examined in thin sections, and the fourth was the published literature. The first collection studied was that of the University of Chicago. Although this included about 50 specimens, it nevertheless clearly represented only specific areas of the United States, and many important sandstone-bearing regions were entirely unrepresented. The Chicago Museum of Natural History provided a second collection. From this collection 30 specimens were obtained; these also represented only certain areas of the United States and Europe. Fortunately, this collection represented areas different from those in the University of Chicago's collection. The third and largest collection was the 205 specimens forwarded to the writer by geologists throughout the United States. The author found most of the descriptions in the literature to be in-

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[1] Manuscript received December 13, 1948.

complete; hence only those definitive enough to indicate clearly the correct classification of the sandstones were used. Microphotographs, wherever included, proved invaluable in interpreting the descriptions. Many of the relatively recent descriptions were decidedly inferior to descriptions dated back thirty to fifty years.

Even if the varied sources used had yielded a sample of adequate geographic and stratigraphic diversity, many limitations would still remain. First, the author assumed that the relative abundance of the several sandstone types could be obtained by count or tabulation of specific formations and lithologic units, which, however, were of varying thickness and extent. In effect, this assigned equal weight to each specimen irrespective of the amount of sandstone that the particular specimen might represent. Thus a specimen of Dakota sandstone (which represents a formation extending over hundreds of thousands of square miles) carries no more significance than a specimen of the Martinez sandstone (which occurs over a limited area). The actual importance of this factor is unknown.

It would be desirable to use actual volume units in measuring the sandstones, but the problem of arbitrarily selecting and collecting sandstone specimens from specific sections throughout the United States and assigning the appropriate weight to each specimen was not feasible.

The collection of a representative sample of sandstones is further complicated because some distinctive formations are easily delineated and their extent and thickness may be much more precisely known than less distinctive formations. In particular, the graywackes tend to be difficult to measure. These formations commonly occur in thin layers from 2 inches to several feet thick, separated by larger layers of shale or slate. In such cases the graywacke may form one-third to one-fourth the total formation, but the entire unit may still be designated a slate. One of the many examples of this is the Tyler Slate of Michigan.[2] Here, although the total thickness of graywacke may be several thousand feet, the individual beds, not mappable or named formations, are commercially of no importance and therefore not collected, described, or analyzed.

Duplication of specimens is another imperfection in the sample. For example, the final sample used in this study included specimens of the Pottsville sandstone from Alabama, Georgia, and Ohio. All three of these specimens were included because each seemed representative of the sandstones in the particular locality in which the specimen was found, and these localities of origin were widely separated. Where more than one specimen was available from the same formation and the original sources were close together, only one specimen was used.

The assembled data included a total of 277 specimens. Of these, 94 were obtained from the literature and 183 from examination of thin sections. These data included sandstones of all geologic ages, as shown in table 1.

CLASSIFICATION OF SANDSTONES

The classification used in this study is based on the mineral composition of the detrital grains, and only those specimens in the accepted size range for sandstones were used ($\frac{1}{16}$–2 mm.). Two methods of tabulation were used. In the first classification the sandstones were separated into three series: (1) orthoquartzite series, (2) arkose series, and (3) graywacke series. Krynine has given the following definitions (1948, p. 149):

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[2] F. J. Pettijohn, personal communication.

> The quartzite series of sediments can be popularly defined as being made up of "clean sands." It contains one detrital (clastic) end-member, consisting almost exclusively of quartz, occurring generally as medium-sized grains. . . .
>
> The graywacke series of sediments can be popularly defined as a "dark dirty sand" or as a "pepper-and-salt sand" for the lighter-colored but strongly cherty varieties.
>
> Graywackes contain two or more major detrital clastic end-members: grains of quartz, chert, slate, schists, phyllites, etc., and a fine-grained matrix made of finely divided particles of micas (illite, sericite, muscovite, biotite, chlorite). . . .

TABLE 1

DISTRIBUTION OF SANDSTONE SPECIMENS WITH RESPECT TO GEOLOGIC AGE

Geologic Age	Observed Specimens	Published Descriptions
Cenozoic	10	3
Cretaceous	14	15
Jurassic	14	6
Triassic	14	8
Permian	5	4
Pennsylvanian	32	17
Mississippian	14	
Devonian	11	5
Silurian	8	7
Ordovician	8	10
Cambrian	18	12
Pre-Cambrian	8	10

> The arkosic series of sediments can be popularly defined as "ashy, light-gray (or red) dirty sands with much feldspathic materials."
>
> The arkose usually contains one or two major detrital clastic end-members, grains derived from the rapid erosion of a granitic detritus), in some cases mixed with a finer grained matrix of clay minerals (kaolinite-montmorillonite type) and frequently much iron oxide.

These definitions, while applicable in a general sense, are not precise enough for the separation of certain intermediate-type sandstones, and accordingly the following definitions were used in this paper. The orthoquartzite series includes sandstones with greater than 90 per cent of the detrital constituents made up of quartz grains; they contain, therefore, less than 10 per cent of clay or argillaceous material and also less than 10 per cent of feldspar. The arkose series includes sandstones with greater than 10 per cent of the detrital consituents made up of feldspar and with less than 10 per cent of clay or argillaceous material. The graywacke series includes sandstones containing greater than 10 per cent of clay or argillaceous material.

A second method of tabulation was used to classify the sandstones into more restricted classes. These classes are defined as follows:

1. An orthoquartzite is a sandstone with greater than 90 per cent of the detrital constituents made up of quartz grains.
2. A feldspathic sandstone is a sandstone with feldspar grains making up 10–25 per cent of the total detrital constituents and with the argillaceous components less than 10 per cent.
3. An arkose is a sandstone with feldspar making up greater than 25 per cent of the detrital constituents and with the argillaceous components less than 10 per cent.
4. A graywacke is a sandstone with the argillaceous component making up greater than 25 per cent of the total rock.
5. A subgraywacke is a sandstone with the argillaceous component making up 10–25 per cent of the total rock, plus those sandstones having less than 10 per cent of feldspar or argillaceous material, as well as less than 90 per cent of quartz.

Although the threefold classification has perhaps a wider usage among geologists, this author believes that it is inferior to the fivefold system. The arkose, graywacke, and orthoquartzite series classification is especially weak in the consideration of the graywacke type. The history of the term "graywacke" from 1808 to the present has been traced by Pettijohn, and he concludes (1944, p. 944) that

> graywacke connotes a type of sandstone marked by (1) large detrital quartz and feldspar ("phenocrysts") set in a (2) prominent to dominant "clay" matrix (and hence absence of infiltration or mineral cement which may on low grade metamorphism (diagenesis) be converted to a chlorite and sericite and partially replaced by carbonate, (3) a dark color, (4) generally tough and well indurated, (5) extreme angularity of the detrital components (microbreccia), (6) presence in smaller or larger quantities of rock fragments, mainly chert, quartzite, slate or phyllite, and (7). . . .

This particular type of sandstone makes up 15 per cent of all the sandstones ex-

TABLE 2

SUMMATION OF THE SANDSTONE TYPES ACCORDING TO GEOLOGIC AGE

GEOLOGIC AGE	OBSERVED SAMPLE					DESCRIPTIVE LITERATURE				
	Orthoquartzites	Arkoses	Feldspathic Sandstones	Graywackes	Subgraywackes	Orthoquartzites	Arkoses	Feldspathic Sandstones	Graywackes	Subgraywackes
Cenozoic	2	2	1	3	3	2	1	....	....	....
Cretaceous	9	....	3	2	2	8	1	....	2	4
Jurassic	5	1	1	3	4	5	....	1	....	....
Triassic	5	3	1	1	3	2	4	2	....	1
Permian	3	....	1	1	2	....	2	1	1	....
Pennsylvanian	12	....	1	4	15	10	1	....	4	2
Mississippian	11	....	1	1	5					
Devonian	4	....	....	3	6	1	....	....	4	....
Silurian	6	....	1	2	....	2	2	....	1	1
Ordovician	5	....	....	2	1	6	....	....	2	2
Cambrian	14	....	4	2	1	5	2	1	....	2
Pre-Cambrian	5	4	....	2	1	2	1	2	5	1
Uncertain	1	2	1	2	3	....	....	....	....	....

TABLE 3

RELATIVE ABUNDANCE OF SANDSTONE TYPES IN PERCENTAGES

SANDSTONE TYPES	OBSERVED SAMPLE		DESCRIPTIVE LITERATURE		TOTAL	
	No.	Percentage	No.	Percentage	No.	Percentage
Orthoquartzites	82	45	43	46	125	45
Feldspathic sandstones	15	8	5	5	20	7
Arkoses	12	7	14	15	26	10
Subgraywackes	46	25	13	14*	59	21*
Graywackes	28	15	19	20	47	17

* The percentage of subgraywackes is probably too low, since many indefinite descriptions of the literature were discarded and, of these, the subgraywackes, being of intermediate type, were probably eliminated disproportionately. The value obtained in the observed sample (25 per cent) is probably more nearly correct.

amined, as shown in table 3. A larger group of sandstones (subgraywacke) has characteristics intermediate between the orthoquartzites and the graywackes. It would seem inaccurate, therefore, to this author, to classify together these two types of sandstones under one general heading (graywacke series) which previously has been defined to include only the smaller group.

TABLE 4

Rock Types	Per Cent
Orthoquartzite	45
Feldspathic sandstones	7
Arkoses	10
Subgraywackes	21
Graywackes	17

No attempt has been made in this paper to consider unusual or atypical sandstone types; and, of the 185 specimens examined, only 5 per cent did not fit into one of the five "types." These specimens were excluded from the study.

The observed specimens as well as those described from the literature have been classified and assembled in table 2. These data are further summarized in table 3. The relative abundance of each of the sandstone types as shown in this summary is given in table 4.

If these data are computed on the basis of a threefold system of classification, the percentages of each of the types are as shown in table 5.

TABLE 5

Series	Per Cent
Orthoquartzite	45
Arkose	17
Graywacke	38

It is quite probable that the values obtained for the orthoquartzites are excessively high because this type has been more extensively collected and described by geologists. The percentage of arkoses and arkose series sandstones are significantly lower than some earlier estimates. Also, as noted in table 3, it is probable that the value shown for the subgraywackes is too low.

The volumetric distribution of the sedimentary rocks (including both conglomerates and sandstones) has also been given by Krynine (1948, p. 156). Considering these apart from siltstones and shales, he finds the percentages of each type shown in table 6.

TABLE 6

	Per Cent
Orthoquartzites	22½
Low-rank graywacke	35
High-rank graywacke	10
Arkose	32½

Krynine also indicates by diagram the average mineral composition of sandstones (p. 151). Measurement of his diagram indicates the average compositions are as shown in table 7. It is important to note that the percentages given by Krynine refer to total composition, i.e., including both detrital and chemical constituents.

TABLE 7

AVERAGE MINERAL COMPOSITION OF COMMON SANDSTONE TYPES IN PERCENTAGES (AFTER KRYNINE)

Minerals	Arkoses	High-Rank Graywacke	Low-Rank Graywacke	Quartzite
Micas and chlorite	5	22	22	0
Quartz	48	46	40	79
Feldspars	25	20	4	3
Clays	8	0	5	0
Carbonate	5	2	7	6
Chalcedony	6	7	21	13

If the figures as obtained from the sample used in this study are applied to sandstones in general, it is likely that the individual percentages would vary considerably, but the general order of magnitude should remain.

Because the definitions of each sandstone type (as given earlier in this paper) allow considerable variation within each group, an infinite number of variations was possible; but, in examining the specimens, an effort was made to select those which were typical or ideal examples of the group in question. For example, according to the definitions used, a graywacke might have either a quartz content or a feldspar content, varying anywhere between 0 and 75 per cent; yet, on the basis of actual inspection, the specimens which most clearly conform to the generalized impression conveyed by the term "graywacke" are sandstones having a quartz content of about 25 per cent and a feldspar content of about 20 per cent (plus about 45 per cent argillaceous material). This ideal example is likely to be near an arithmetic mean of all sandstones in this particular type.

TABLE 8

MINERAL COMPOSITION OF SANDSTONE TYPES IN PERCENTAGES

MINERALS	ORTHO-QUARTZITES*		FELDSPATHIC SANDSTONES†		ARKOSES‡		SUBGRAY-WACKES§		GRAY-WACKES‖	
	Average	Range	Average	Range	Average	Range	Average	Range	Average	Range
Quartz	95	90–99	77	65–85	58	40–70	75	65–83	25	15–33
Feldspar	2	0–5	13	10–15	27	25–30	7	0–8	18	2–30
Rock fragments	1	0–4	3	0–10	3	0–10	5	0–6	7	1–25
Chert	2	0–5	5	0–10	5	0–10	5	1–10	4	1–10
Argillaceous material	1	0–4	2	0–5	5	0–10	10	2–20	46	20–60

* Includes specimens of the following sandstones: Baraboo, Bass, Cheyenne, Chinle, Graydon, Hepler, Hinkley, Ionian, Jelm, McElmo, Nugget, and Palestine.

† Includes specimens of the following sandstones: Frontier, Goodenough, Hector, Ireland, Kome, LaMotte, and Unkpapa.

‡ Includes specimens of the following sandstones: Hector, Torridonian, and arkosic sandstones from Mount Tom, Massachusetts, Springfield, Massachusetts, and Maryland.

§ Includes specimens of the following sandstones: Bighorn, Dakota, Oswego, Sundance, Third Bradford Sand, and a Cambrian sandstone from the Lake Superior region.

‖ Includes specimens of the following sandstones: Bayard, Catskill, Chico, Franciscan, Monterey, Rensselaer Grit, Upper Slate Member (of Wisconsin), and an Ordovician graywacke from Scotland.

Forty specimens, each of which appeared to be a good example of one of the sandstone types, were examined, and the average mineral composition of each type is shown in table 8. The range of each of the more common minerals is also included. It is important to note that the data of table 8 do not include the introduced mineral cements of the sandstones. However, in the case of the graywackes and any of the subgraywackes it is impossible to exclude any particular portion of the rocks on the basis of secondary introduction. Therefore, the entire rock must be included in the measurement of the mineral composition of many of the specimens examined. Utilizing the data of table 7 and table 8, it is possible to estimate the mineral composition of an average sandstone. As shown in table 9, the quartz and feldspar content of the average sandstone has been estimated to be 74 and 8 per cent, respectively.

Using entirely different methods, Clarke has estimated the average quartz and feldspar composition of sandstones to be 66.8 and 11.5 per cent, respectively (1924, p. 33). The general agreement between the values as computed in this paper and those of Clarke indicates that the order of magnitude of the values derived in this paper is correct.

TABLE 9

COMPUTATION OF QUARTZ AND FELDSPAR CONTENT OF THE AVERAGE SANDSTONE

Types (1)	Percentage of All Sandstones (2)	Percentage of Feldspar (3)	Col. (2)× Col. (3) (4)	Percentage of Quartz (5)	Col. (2)× Col. (5) (6)
Orthoquartzites	45	2	0.0090	95	0.4275
Feldspathic sandstones	7	13	0.0091	77	0.0539
Arkoses	10	27	0.0270	58	0.0580
Subgraywackes	21	7	0.0147	75	0.1575
Graywackes	17	18	0.0306	25	0.0425
Totals	....	....	0.0804 or 8 per cent	....	0.7394 or 74 per cent

TABLE 10

RELATIONS OF DOMINANT CEMENTS AND THE GEOLOGIC AGE OF SANDSTONES

GEOLOGIC AGE	DESCRIPTIVE LITERATURE		OBSERVED SPECIMENS		LITERATURE AND OBSERVED SPECIMENS	
	Silica Cement	Carbonate Cement	Silica Cement	Carbonate Cement	Silica Cement	Carbonate Cement
Cenozoic	2	....	1	2	3	2
Cretaceous	2	5	5	6	7	11
Jurassic	2	4	6	2	8	6
Triassic	1	....	5	3	6	3
Total	7	9	17	13	24	22
Permian	1	....	3	1	4	1
Pennsylvanian	6	3	18	7	32	11
Mississippian			8	1		
Devonian	1	....	4	2	5	2
Total	8	3	33	11	41	14
Silurian	4	....	9	2	13	2
Ordovician	4	....	4	1	8	1
Cambrian	3	....	15	3	18	3
Pre-Cambrian	2	....	6	2	8	2
Total	13	....	34	8	47	8

SANDSTONE CEMENTS

The second main objective of this study was the determination of the cementing minerals of the sandstones, with special reference to the ratio of the silica-cemented sandstones to those cemented by carbonate. In cases where two or more different cementing minerals were present, the specimen was disregarded unless one mineral was clearly dominant. Both the rocks described in the literature and the available specimens were classified as to cement. Table 10 presents these data, separated into the geological time units. The variations in the percentages of silica and carbonate cement are summarized in table 11.

The definite increase in the percentage of carbonate-cemented sandstones in the later geologic periods is very evident. This increase may be due to any one of a multitude of factors concerning the deposition and subsequent history of the sandstones, or it may possibly be due to an inherent preference of one particular sandstone type for a particular cement. For example, if the orthoquartzite type sandstone should be present in a larger percentage in the older geologic periods and if this particular type should be predominantly associated with silica cement, this might explain the variation which is found in the cements with respect to age. In an effort to test this possibility we may consider the orthoquartzite type (which has been shown to be the largest single type) more closely.

TABLE 11

Period	Per Cent Silica	Per Cent Carbonate
Triassic to Recent	52	48
Devonian to Permian	75	25
Pre-Cambrian to Silurian	84	16

First, tables 1, 2, and 3 show that there is no significant variation in the percentage of orthoquartzites with respect to time. The values of table 3 can be summarized as shown in table 12.

TABLE 12

RELATIVE ABUNDANCE OF ORTHOQUARTZITES WITH RESPECT TO GEOLOGIC AGE

Age	Total No.	No. of Orthoquartzites	Per Cent of Orthoquartzites
Triassic to Recent	81	35	47
Devonian to Permian	108	45	42
Pre-Cambrian to Silurian	92	48	52

It can also be shown that there is no variation in any of the other sandstone types of sufficient magnitude to be responsible for the variation in mineral cements which does exist. Furthermore, table 10 shows that 28 per cent of all sandstones examined having a mineral cement are cemented with carbonate. Of a total of 63 orthoquartzites examined, 14, or 22 per cent, showed a carbonate cement. There does not appear to be any significant discrepancy between these two values. Therefore, we may assume that the variation in mineral cements is due not to a variation in the relative amount of each type of sandstone but rather to some physical process present in the history of the sandstone. Just why the carbonate cement is more abundant in the later geologic periods is a question yet unanswered.

This can be further demonstrated by an examination of a sample consisting entirely of one type of sandstone. Table 13 shows the relative abundance of silica and carbonate cements in a group of 77 orthoquartzites, and these data are summarized in table 14.

The values thus summarized from table 13 show marked agreement with those summarized from table 10, and therefore it is again demonstrated that the variation in cement among sandstone studies is independent of the particular type of sandstone.

TABLE 13

DISTRIBUTION OF SILICA AND CARBONATE-CEMENTED ORTHOQUARTZITES WITH RESPECT TO GEOLOGIC AGE

Period	No. with Silica Cement	No. with Carbonate Cement
Cenozoic	1	1
Cretaceous	6	2
Jurassic	1	3
Triassic	3	2
Permian	1	1
Pennsylvanian	10	2
Mississippian	10	2
Devonian	2	1
Silurian	4	2
Ordovician	4	1
Cambrian	13	1
Pre-Cambrian	4	0

An interesting observation was made on the carbonate-cemented sandstones. Of 44 different specimens examined, almost every one showed some evidence of replacement of detrital quartz grains by carbonate. This process was seen in all stages of development, varying from slight corrosion of some quartz grains to virtually complete replacement of others.

TABLE 14

Period	Per Cent of Silica	Per Cent of Carbonate
Triassic to Recent	63	37
Devonian to Permian	79	21
Pre-Cambrian to Silurian	86	14

On the basis of the few samples observed, it would appear that this process of replacement or partial replacement is almost universal in the carbonate-bearing sandstones. Should this replacement process continue for a long enough period, it should be possible for a sandstone layer to be completely converted to a limestone. The writer is not aware of any verification of this completed conversion.

CONCLUSIONS

The following observations and conclusions have been made, based on the examination of the specimens used in this study:

1. On the basis of the detrital mineral components, sandstones may be conveniently separated into five classes, and the relative abundance of these classes is as follows:

	Per Cent
A. Orthoquartzites	45
B. Feldspathic sandstones	7
C. Arkoses	10
D. Subgraywackes	21[3]
E. Graywackes	17

2. A comparison of the carbonate-cemented sandstones with the silica-cemented varieties shows a definite increase in carbonate cement in the later geologic periods.
3. Almost every sandstone specimen showing carbonate cement also displayed partial replacement of the detrital grains by the carbonate.

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[3] As stated in table 7, this value is probably too low, and a better estimation might be 25 per cent.

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RECENT RESULTS OF INVESTIGATIONS ON THE FELDSPARS[1]

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ABSTRACT

Careful work with the universal stage on twinned plagioclases has shown that certain optical properties are dependent upon the thermal history of the feldspar. The generally accepted determinative curves for the plagioclases were drawn from feldspars in plutonic rocks and crystalline schists; the volcanic rock plagioclases do not fit these curves. New curves, for high-temperature plagioclases, have been constructed, but are not published here. A discussion of the relation of crystal habit and optical properties of potash feldspars to their genesis is also presented.

No other group of minerals is of so much interest to mineralogists and petrologists as are the feldspars. The feldspars are of such importance as components of the earth's crust that a proper understanding of them is essential for the genesis and classification of rocks. Although scientists of many countries have worked with these minerals, very much more intensive investigation is needed.

This article embodies some recent results bearing on rock genesis and a topic of special importance to petrologists and geologists, namely, the exact determination, with a sensitive and simple method, of the percentage of anorthite in plagioclase as a function of the temperature of crystallization. It will be shown how crystallographic and optical data which seem to be unimportant can be used to interpret the genesis of feldspars.

The first aim in any investigation of plagioclases is the determination of the percentage of An as precisely as possible. Methods for this purpose have been constantly improved. It is possible to get results by using extinction angles in oriented thin sections by means of the "classical methods"; but it is possible to do so only if one takes into consideration the differences, to be discussed later, between plagioclases in plutonic and metamorphic rocks and those in volcanic rocks. The Fedorow universal stage should, however, always be used. No particular technique can be insisted upon here; universal-stage methods are not "universal," and they cannot be used for the investigation of very small crystals, on which extinction methods may be useful. Determinations with the aid of refractive indices, a method which has been elaborated and improved in the United States, shows some factor of uncertainty if one does not watch for the effects to be described later. It is best to use all known methods of determination, checking one against another. One should use the best available method for the material being investigated.

A criticism of the extinction method arises from the fact that plagioclases are not always two-component systems but contain at least three components, the orthoclase content being up to 6 mol per cent. Possibly variation in the amount of orthoclase changes the optical properties and thereby creates an uncertainty in the use of extinction curves. For this reason the extinction methods became discredited, and a preference for the universal-stage method arose. This, however, was a mistake, as will be shown. (It was also

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thought that carnegieite [$NaAlSiO_4$] might be present in solid solution [Washington and Wright, 1910; Barth, 1930]. This should also discredit the use of extinction curves, but later investigations [Dittler and Köhler, 1932; Ernst and Nieland, 1934; and others] showed that practically no carnegieite is taken in the feldspars, although results do not appear to be final on this matter [see, e.g., Lundegardh, 1941].)

If one determines the position of the indicatrix with the aid of the universal stage and relates it to a measured trace (of cleavage or twin plane), as was done by M. Reinhard (1931), M. Berek (1924), and V. Nikitin (1936), it will commonly be noted that the traces of the planes deviate from the position theoretically expected. The determined pole of M (010), in particular, at times deviates excessively from the migration curve holding for M, for example, in the diagram, table 2, by M. Reinhard (see, e.g., the determinations by G. Paliuc [1932] and others). It was earlier thought that this deviation was due to the potash component and that from this deviation the amount of orthoclase could perhaps be at least estimated (Reinhard, 1931).

There was, at first, no explanation for these deviations. In order to check the above doubtful explanation, the author attempted another approach, and the method is as follows: One measures the indicatrixes of both twins with the universal stage and plots the positions of the main indices, α, β, γ, and the measured optic axes on the stereographic projection (see the procedure of Reinhard). It is also possible to measure or construct the pole of M (it lies on the intersection of the great circles $\alpha\alpha$, $\beta\beta$, and $\gamma\gamma$). Because the optical directions can be measured more exactly than can the crystallographic ones, the construction of the pole should be preferred. In such a diagram it is possible to measure several angles, at least three, namely $\alpha\alpha'$, $\beta\beta'$, and $\gamma\gamma'$ and perhaps some angles between equivalent or inequivalent optical axes of the twins, and to use them for the determination of An content. In the case of the albite law, for example, the angle between the optic axis A of the one individual and the axis B of the other can be used very successfully (for more detail see Köhler, 1941*a*, *b*).

The extinction methods give only one value that can be used; the universal-stage method also gives the position of but one trace of a plane (rarely two if it is possible to measure poles M and P). However, our kind of determination leads to the measurement of several angles, all of which should give the same An content if the measurements are correct. The several measurements are mutually consistent and therefore reliable, which is a distinct advance. It is, of course, necessary first to construct the angles between α, β, and γ and the optical axes of the two different twins for the known twin laws, using the orientations of "standard plagioclases." This is easily done. In this way one gets for the pericline law the angle $\alpha\alpha^{\pi}$, $\beta\beta^{\pi}$, $\gamma\gamma^{\pi}$, and AB^{π}, BB^{π}; for the Carlsbad law the angle $\alpha_1\alpha_2$, $\beta_1\beta_2$, $\gamma_1\gamma_2$, A_1A_2, B_1B_2, etc. The same values were used by F. Becke (1906) in his "konoskopischen Zeichentisch-Methoden"; however, he did not use the universal stage and was therefore confined to a few section orientations. The universal stage makes us independent of any specific orientation and enables us to read all necessary angles immediately in the plotted stereogram (see the curves of such angles in Köhler, 1941*a*).

So much may be said in the way of preliminary information. If one now measures, in the above way, a labradorite with albite twinning from a gabbro and if one plots the determined values of α, β, and γ and the optic axes in the stereographic projection (method of Reinhard) and if one plots in the same way the measured position of the pole of *M*, it is then possible to read from the stereogram the following angles (see Köhler, 1941*b*):

$$\alpha\alpha' = 159°, \quad \beta\beta' = 48.5°, \quad \gamma\gamma' = 54°,$$

$$AB' = 44.5°, \quad BB' = 108°.$$

These five angles give, respectively, an anorthite content of

$$50, \quad 51, \quad 49.5, \quad 51, \quad 50.5 \text{ per cent}.$$

Using the position of the pole *M*, one gets in both twins 51 per cent An. The agreement is very good, although admittedly this case was favorable because the twin lamellae were broad and easily measured. The pole of *M* fits Reinhard's migration curve. If the individuals are not easily measured (and assuming careful work by the microscopist), one will get deviations up to approximately 3° above and below the average. If a measurement is made on an untwinned crystal and one measures a wrong position of the pole of *M*, it would be possible to fit the pole on the migration curve and have it show a value for the An content which appears correct but is not necessarily so. It is important and can be immediately seen in the above example that the albite law holds because the great circle of the pole *M* is symmetrical to the traces of the axial planes.

If one examines a number of plagioclases of plutonic rocks and crystalline schists in this way, then one will see that within the limit of error the pole of *M* always lies on the migration curve and the values of the angles all show the same amount of anorthite. However, if one measures a plagioclase of a volcanic rock, quite different relations are found. For instance, using labradorite from an andesite, one will find the following angles and corresponding An contents which, in the first example, are taken from the curves used for the values given above:

$$\alpha\alpha' = 158° = 50 \text{ per cent An},$$

$$\beta\beta' = 62° = ? \text{ An},$$

$$\gamma\gamma' = 67° = 61 \text{ per cent An};$$

$$AB' = 52° = 58 \text{ per cent An},$$

$$BB' = 109° = 56 \text{ per cent An}.$$

It can be seen that the value of the anorthite content varies between 50 and 61 per cent! The high value of 62° for $\beta\beta'$ does not fit on the curve at all. It is therefore impossible to determine the labradorite with this curve. It is not possible to explain this fact as a result of measurements that are in error, since the discrepancy is inherent in the fundamental properties of the feldspar. It is peculiar that in this case the pole of *M* lies far to the right side of the standard migration curve. Neither can this discrepancy be explained by assuming orthoclase in solid solution; one cannot comprehend why such an amount of orthoclase should be observed only in volcanic rocks. It should be mentioned, moreover, that we have chemical analyses of plagioclases from plutonic rocks with 5–6 mol per cent orthoclase in which the anorthite content is well determinable and the *M*-poles of which lie on the normal migration curve. *The reason therefore seems to be that the optics of volcanic rock-plagioclases (position of the indicatrix) differ fundamentally from those of plutonic rocks.* The high temperature of crystallization of plagioclases in volcanic rocks produces optics which differ measurably from those of plagioclases of the same chemical composition from crystalline schists or plutonic rocks which are crystallized at lower temperature. As concerns the latter, it can be said without any doubt (as shown below) that the original high-temperature optics changed slowly to the low-temperature optics.

When we know this, the reason for deviation is clarified. It is strange, particularly in the light of our knowledge of the potash feldspars (sanidine-orthoclase-microcline), that the temperature dependency of the optics was not discovered until the sensitive twin optics (Köhler, 1941*a*, *b*) were used. Furthermore, it was due to chance that all plagioclases used for the construction of all the curves were taken from plutonic rocks. Our extinction curves were therefore useless for the determination of volcanic rock plagioclases. It is necessary to establish new curves for such determinations. We have but a few values of optical orientation for such plagioclases (Ernst and Nieland, 1934); we must therefore look for convenient methods of determination. It was thought plausible that plagioclases formed synthetically from a melt should show high-temperature optics. Mixtures were therefore made with 70, 80, 90, and 100 per cent anorthite, and the optical properties of these crystallized synthetic products were measured with the universal stage (see the work of H. Tertsch, 1941*a*, *b*). Another experiment has shown that the position of the indicatrix of low-temperature plagioclases (which were measured before the experiment) was changed to a position measurably different by heating. Tom. F. W. Barth (1931) found incontestable evidence of such a change, which, however, was unjustifiably questioned by C. T. Barber (1936). H. Scholler (1941) studied the temperature dependency of optics in basic and intermediate members in an excellent way. H. Tertsch (1944–45) made a similar study with acid plagioclases, the synthesis of which from the melt is not possible. In this way all curves of the twin optics and all extinction curves for the high-temperature range could be found. These curves are not so precise as the low-temperature curves, but they are exact enough to determine the anorthite content satisfactorily. Turning again to the earlier considered labradorite from the andesite, we find the following values, using the new data:

$\alpha\alpha' = 158° = 51\text{-}52$ per cent An

$\beta\beta' = 62° = 54$ per cent An .

$\gamma\gamma' = 67° = 51.5$ per cent An ,

$AB' = 52° = 50$ per cent An ,

$BB' = 109° = 52$ per cent An .

This agreement is excellent and is not obtainable with the "low-temperature" curves.

With respect to most plagioclases (andesine to labradorite), the differences between the two ranges of temperature are fairly large and are always clearly distinguishable. It is a disappointing fact, however, that all curves cross at about 80–90 per cent anorthite and that at the albitic end differences are slight. Not many cases may be found, however, in which the distinction cannot be made, as optical measurement of the twins gives several figures, one or the other of which may lead to a decision. For further information see Köhler (1941*a*, *b*) and Tertsch (1941*a*, *b*). The new curves have not as yet been published in any comprehensive form; this will be done in a monograph on the feldspars that the author has ready for publication in the Verlag Springer, Heidelberg, Germany.

What are the advantages of the new type of determination? The more precise nature of mutual checking of angles has already been mentioned. Also, the microscopist, working without the universal stage and using extinction curves, has to use the newly constructed curves because he obtains a value of the anorthite content in the plagioclases of volcanic rocks that is too high by 10–15 per cent through the use of the old curves. An error of this magnitude is unpardonable for the modern petrologist. In no case, therefore, should such plagioclases be determined by other methods, especially not by means of extinction angles. Synthetic plagioclases from the melt can be determined only by means of high-temperature curves; all previous data that indicate that a synthetic pure anorthite, measured optically, shows 100 per cent An are wrong, as all the older methods, when used precisely, really show in this case only 90–92 per cent An.

Another advantage is that one can determine whether the relic in a crystalline schist was originally from a volcanic rock or not (see examples in Köhler, 1941*a*, *b*). We can therefore describe the history of a rock more precisely. It is to be expected, however, that further improvements will be possible that will give us additional information on the past history of plagioclases (reheating in contacts and inclusions, for instance). The author would be grateful for information on measurements of such plagioclases. It would also be interesting to measure the plagioclases of meteorites, something that the author has had no opportunity to do.

When one heats plagioclases of plutonic rocks as mentioned above, one is struck by the fact that they may be brought to the high-temperature form very easily. Crystals that had once been in the high-temperature state are very easily returned to that state; it seems that "relic" structures of the high-temperature form initiate and speed up this process. Material that crystallized at low temperature is very difficult to transform to the high form. The observations of H. Tertsch (1944–45), who found no change in albites from Alpine fissures, with temperatures just below the melting point, but who did find measurable change in albite from Mariupolit treated in the same way, are explainable on the basis of this different behavior.

Similarly important, but until now not so easily described and not so useful in determining genesis as are the results of the research on plagioclases mentioned above, are observations on potassium feldspars, which should be given more consideration than they have received. The type of microcline formed and the type of perthite exsolution that shows different values of the axial angles $2V\alpha$ give some indication as to the history of the development of the feldspars, and therefore some conclusions can be drawn concerning the history of the rocks, especially the crystalline schists. The crystal habit and the type of twinning is also important; it has been observed after long experience that crystals with elongated X-axes or (more rarely) tabular after M (010), with the main faces (001), (010), (110), and ($\bar{2}$01) and with subsidiary faces (130), (021), and ($\bar{1}$11), are characteristic of high-temperature formation, possible only in melts. Pegmatitic K-feldspars and adularias, on the other hand, show another habit. At high crystallization temperatures the Carlsbad law is most commonly found, in pegmatites the Baveno law, and adularias for the most part show the Manebach law (associated with the Baveno law), but no Carlsbad twinning. Especially interesting are the Carlsbad twins which overlap and therefore grow somewhat irregularly; the existence of the ($\bar{2}01$) plane is a necessary condition, however (see Tertsch, 1936; Köhler and Raaz, 1947). Such twins have the (010) plane as the contact plane only in the core, which means that more or less tabular single crystals simultaneously growing in an oriented position became twinned when they were still enclosed in a liquid. When one finds such relations in, for example, granitic gneisses, in which large crystals of this kind quite often are to be found, one can justifiably conclude that these crystals had previously been granite feldspars. When such K-feldspars are twinned according to this law in pegmatites, however, they grow with (010) as the contact plane, and overlapping is not seen because the ($\bar{2}01$) plane is lacking, the ($\bar{1}01$) plane being present instead. On the other hand, such "Augen" have often grown in crystalline schists by later infiltration of alkali and silica, which means that it is not necessary to consider them as deformed relics. Hints as to genesis are supplied by such crystallographic observations, but more needs to be done to clarify the issue; for example, the type of microcline cross grating should be considered. This grating depends essentially on the temperature conditions after crystallization and on the tectonic history of the potash feldspar. The amount and kind of perthitic exsolution is connected with all this, and hand in hand with this exsolution can be seen a slow increase in the axial angle $2V\alpha$.

Potash feldspars that have been at high temperatures and rapidly cooled retain the soda content more or less isomorphously, microcline grating is not visible, and the axial angle remains small. With slower cooling, perthitic exsolution appears, and the grating is obvious (see granite feldspars). The size of $2V\alpha$ ranges between 65° and 75°. The potash feldspars that originally certainly crystallized in the monoclinic form (Köhler, 1948) change to a triclinic modification; this change, being aided by tectonic forces which, with increased intensity, increases the "coarseness" of the exsolution of albite and produces microcline grating. It is always quite obvious that the grating is most distinct in the neighborhood of the broad albite "stringers," as in these places the exsolution of the albite destroys the originally monoclinic lattice, and the newly crystallized K-feldspar assumes the triclinic modification which is now the stable one. The crystal, by twinning polysynthetically according to the albite and pericline laws, attempts to retain its past monoclinic habit (Köhler and Raaz, 1947).

When the original rock (granite, for example) is very strongly changed to a crystalline schist (granite gneiss), all the albite will be exsolved and will form independent albite crystals in the gneiss, and the K-feldspar nearly free of soda will have very coarse grating structure. In addition, the axial angle (around α) of the K-feldspar will be $>80°$ and, in fact, $>90°$ more often than has been realized (optically positive "isomicrocline").

It would therefore be possible, using the phenomena mentioned above, to obtain some insight into magmatic and crystalloblastic formation if it were not for the fact that pegmatitic liquids and solutions and formation of feldspars in deep zones of mountain roots cloud the picture. The perthitic K-feldspar from a pegmatite looks very much like strongly exsolved granite feldspar, as regards grating and exsolution structures. If no Carlsbad twinning is present to give one a hint as to the origin of the feldspar, wrong conclusions might be drawn. In such cases field observations may tell more than microscopic investigation of the K-feldspars, and it is necessary to use observation on the plagioclases. If the typical polysynthetic albite twins in the plagioclases are lacking, we have an indication that the crystals were formed at low temperature (for example, granite gneisses of the upper Katazone, such as in the central Alps). If there are multiple repetitions of the lamellae, it is suggestive of high temperature at an earlier time or that the plagioclases came from pegmatites. It can be said without doubt that the multiple repetition of albite lamellae is not a primary result of crystallization but is secondarily induced in the plagioclase. We (Köhler and Raaz, 1947) think that the modification of the plagioclases formed originally from hot melts was a monoclinic one or very nearly monoclinic, or it may be closer to being monoclinic than the plagioclases formed at low temperatures. As K-feldspar with lowering of the temperature is transformed to a finely twinned arrangement, so plagioclases may show a similar effect when the temperature is lowered and mechanical effects come into play. I have tried (1948) to stimulate petrogenetic evaluation of these observations, but the subject still appears somewhat delicate. In a paper by C. Exner (1949) it is shown that the interplay of the influences that change K-feldspar *and* plagioclase gives us some interesting insight into the whole problem.

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DIE HÄUFIGKEIT DER EDELGASE AUF DER ERDE UND IM KOSMOS[1]

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ENGLISH SUMMARY

The cosmic abundance of the rare gases is now known with sufficient accuracy to allow a quantitative comparison with the corresponding data for the earth's atmosphere. It seems safe to assume that the cosmic abundance of neon is closely approximated by Unsöld's value forthe neon content of the Bo star, τ Scorpii. The values for the other rare gases have been determined in a way previously described by the author, viz., by interpolation of the relative values for the different nuclear species in "smoothed" abundance curves. These values, as well as the neon value, may be considered to be correct within a factor of 2. Argon 40 and the two helium isotopes are omitted from consideration because their atmospheric concentration is of radiogenic origin. The comparison of these values with those for the composition of the earth and its atmosphere show that the relative abundance of xenon (the ratio of the number of xenon atoms to the number of silicon atoms) is about 10^7 times smaller on the earth than in the universe, whereas for neon this figure exceeds 10^{11}. A separation of neon from xenon, shifting the ratio of concentrations by a factor of more than 10^4, has obviously taken place during the process of the atmosphere's evolution. This cannot be explained by chemical absorption or solution processes but can be understood by assuming selective diffusion from the earth's gravitational field during a limited epoch of evolution.

In greater detail, it is found that the ratios of terrestrial and cosmic abundances of Ne, A^{36} plus A^{38}, Kr, and Xe, plotted against atomic weight, lie close to a curve, which is given by the following equation:

$$-\log_{10}\frac{N_{\text{ter}}}{N_{\text{sol}}} = 10 \times e^{-0.045M/m_1} + 7.1 \ ,$$

in which N_{ter} = atoms of rare gas present in the earth's atmosphere per 100 atoms of Si present in the earth; N_{sol} = atoms of rare gas in the universe per 100 atoms of Si; M = atom weight of rare gas; and m_1 = atomic mass unit.

The form of this function corresponds to what one would expect to result from selective diffusion. Further discussion of the equation shows that the remarkably small value of the exponent makes it impossible to explain the selective escape from the earth's field of gravitation as a consequence of high temperature only, but that the small value can be understood only if the short period of the earth's rotation at the time of evolution is taken into account. The equation indicates that approximately 10^7 parts to 1 of the rare gases originally present may have escaped from terrestrial matter without undergoing separation, leaving only the small residual fraction to be involved in the separation process.

Attention is drawn to the fact that, according to the given equation, a marked difference should exist between the isotopic composition of atmospheric and that of cosmic rare gases. An investigation of the isotopic composition of Ne and Ar obtained from meteorites, rocks, or natural gases containing an excessive amount of neon might lead to a final decision on the correctness of the assumption suggested.

In den letzten Jahren ist, vor allem durch die grundlegenden Arbeiten des Mineralogen V. M. Goldschmidt (1938) und des Astronomen A. Unsöld (1944) die alte Auffassung in überzeugender Weise bestätigt worden dass sich die Erde und die anderen Planeten, ebenso wie die Meteorite aus einer gasförmigen Urmaterie gebildet haben, die dieselbe chemische Zusammensetzung besass, wie heute die Sonne und auch die Mehrzahl der anderen Fixsterne. Der Verfasser selbst hat (1946) eine Reihe von Argumenten geltend machen können, durch die gezeigt wurde, dass diese überraschend gleichförmige chemische Zusammensetzung der kosmischen Materie im Wesentlichen einer Häufigkeitsverteilung entspricht, mit der sich die Elemente bei einem Entstehungsprozess gebildet haben.

Das Mengenverhältnis aller jener Elemente, die bei einer Temperatur von etwa 2000° K zur Kondensation gelangen, scheint in der Gesamtmenge der

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[1] Manuscript received February 23, 1949.

Erde und in der Gesamtmenge der Meteoriten mit dem auf der Sonne in engen Grenzen übereinzustimmen; diejenigen Elemente jedoch, die bei der genannten Temperatur flüchtig sind oder in einem entsprechenden Medium flüchtige Verbindungen bilden, wie Wasserstoff, Kohlenstoff und Stickstoff, sowie die Edelgase, sind mengenmässig auf der Erde und in den Meteoriten zu einem wesentlich geringeren Anteil vertreten, als auf der Sonne. Es ist klar, dass diese Abtrennung der flüchtigen Anteile der Urmaterie stattgefunden haben muss, nachdem der überwiegende Teil der Erdmaterie bereits in kondensiertem Zustand vorhanden war, denn ein Prozess, der in der Gasphase etwa zu einer praktisch quantitativen Trennung des Xenon vom Jod oder Cäsium geführt haben könnte, ist völlig undenkbar.

TABELLE 1

DEKADISCHE LOGARITHMEN DER HÄUFIGKEITEN SOWIE DES HÄUFIGKEITSVERHÄLTNISSES DER NICHTRADIOGENEN EDELGASE IN DER ERDATMOSPHÄRE UND IM KOSMOS*

(1) El.	(2) M	(3) Vol.-% der Atmosphäre	(4) $\log H_{ter}$	(5) $\log H_{sol}$	(6) $\log \frac{H_{ter}}{H_{sol}}$	(7) Vorhandene % d. Restm.
Ne.	20,2	0,00161	− 8,00	+3,26	−11,26	0,007
Ar.	36,3	0,00345	− 7,67	+1,20	− 8,87	2,1
Kr.	83,8	0,000108	− 9,17	−1,82	− 7,35	55
X.	131,3	0,000008	−10,31	−3,17	− 7,14	91

* Die Häufigkeitswerte sind bezogen auf die Gesamthäufigkeit des Si, die zur Normierung gleich 100 gesetzt ist.

In den letzten Jahren ist es gelungen, verhältnismässig präzise Angaben über die kosmischen Edelgashäufigkeiten zu machen (vergl. Suess, 1949). Diese Fortschritte gaben die Veranlassung dafür, das bereits von Russell und Menzel (1933) eingehend behandelte Problem der irdischen Edelgashäufigkeiten neu aufzugreifen. Auch besteht heute im Wesentlichen Klarheit über den Ablauf der Kernprozesse, durch die die beiden Heliumisotope He^3 und He^4, sowie das Argonisotop A^{40} laufend nachgebildet werden. Der Vorgang der Bildung von He^3 wurde von Libby (1946) diskutiert; der experimentelle Nachweis für die bereits früher (Weizsäcker, 1937; vergl. auch z B. Bleuler u. Gabriel, 1947; Graf, 1948; Suess, 1948) vermutete Bildung von A^{40} durch K-Umwandlung aus K^{40} ist Aldrich und Nier (1948) durch massenspektroskopische Untersuchung von Argon aus kaliumreichen Gesteinen gelungen. Helium, und zwar beide Isotope wandern ständig aus der Erdatmosphäre ab, denn die Atmosphäre enthält nach Goldschmidt (1938) nur etwa 1 % des He^4, das in den letzten $2 \cdot 10^9$ Jahren radiogen aus Thorium und Uran gebildet und aus dem Eruptivgestein im Kreislauf der Verwitterungsvorgänge in Freiheit gesetzt wurde.

Die in Tab. 1 angegebenen Werte beziehen sich auf die anderen, nichtradiogenen Edelgase, die Werte für Argon somit nur auf die Isotope der Masse 36 und 38, denn der ursprüngliche Anteil von A^{40} im kosmischen Argon ist sicherlich kleiner als 1 %, wie eine Abschätzung mit Hilfe der Häufigkeitsregeln (Suess, 1946) zeigt. Für Neon ist die kosmische

Häufigkeit von Unsöld (1941) durch quantitative Spektralanalyse des Bo-Sternes τ Scorpii direkt bestimmt worden. Dieser Stern besitzt, wie Unsöld durch Vergleich der Häufigkeitswerte von sechs anderen Elementen zeigen konnte, sicher dieselbe chemische Zusammensetzung, wie die Sonne. Für die übrigen Edelgase sind vom Verfasser (1949) Häufigkeitswerte angegeben worden, die durch Anwendung von Regeln für die kosmischen Kernhäufigkeiten nach einem Interpolationsverfahren gewonnen worden waren. Bei diesem Verfahren werden zur Interpolation die Häufigkeitswerte der einzelnen Isotopen gesondert herangezogen und Werte erhalten, deren Fehler kaum einen Faktor 2 erreichen dürfte. Die Logarithmen dieser Werte, wie üblich bezogen auf die kosmische Häufigkeit von Silizium gleich 100, sind in Spalte 5 der Tabelle angegeben. In Spalte 3 ist der Gehalt der atmosphärischen Luft an Edelgasen in Volumprozent, in Spalte 4 der Logarithmus der Gesamtmenge Edelgas in der Luft in Atomprozent der Gesamtmenge Silizium im Erdball angegeben, wobei die Siliziummenge zu 15% der Erdmasse angenommen wurde. In Spalte 6 sind die Differenzen der Werte von Spalte 4 und 5, d.h. die Logarithmen der Bruchteile der Edelgasmengen angegeben, die in der Atmosphäre zurückgeblieben sind. Diese Werte sind in Abb. 1 als Funktion des Atomgewichtes dargestellt.

Dass der Edelgasgehalt der Erdatmosphäre ein gutes Mass für die relativen Mengen der Edelgase an der Erdoberfläche ist, lässt sich durch eine genaue Diskussion der Löslichkeiten und Absorbtionskoeffizienten zeigen. Man sieht dann, dass selbst von dem am leichtesten löslichen Xenon nur wenige Prozent der in der Atmosphäre enthaltenen Menge in den Ozeanen gelöst sind, und dass auch die Adsorbtion an den Schneemengen der Polkappen oder an trockenen Wüstensanden gegenüber der Menge freien Xenons nicht ins Gewicht fällt.[2]

Die Zahlenwerte der Spalte 6 bringen zum Ausdruck, dass das Xenon relativ zum Silizium auf der Erde um rund 7 Zehnerpotenzen, Neon jedoch um über 11 Zehnerpotenzen seltener ist, als auf der Sonne. Es ist somit, im Verhältnis, mehr als 10 000 mal soviel Xenon auf der Erde verblieben, als Neon. Dieser Befund lässt sofort erkennen, dass die Edelgase in ihrem gegenwärtigen Mengenverhältnis nicht durch vulkanische Exhalationen oder aus verwitternden Gesteinen abgegeben worden sind, denn Lösungs- oder Adsorbtionsvorgänge können selbst bei so niederen Temperaturen, wie sie heute auf der Erdoberfläche herrschen, eine Trennung des Neon vom Xenon höchstens um etwa einen Faktor 30 hervorrufen. So beträgt z.B. das Verhältnis der Löslichkeiten von Xenon und Neon in Wasser bei 0° C: 11, 4, bei 50° C: 5, 6 (S. Valentiner, 1927). Die Löslichkeit dieser Edelgase im magmatischen Schmelzfluss dürfte sich wegen der hohen Temperatur kaum um mehr als einen Faktor 2 unterscheiden. Es ist völlig ausgeschlossen, dass Vorgänge einer selektiven Adsorbtion oder Lösung einen Trennfaktor der Grössenordnung 10^4 hervorrufen können.

Das gegenwärtige Mengenverhältnis der Edelgase in der Erdatmosphäre lässt sich kaum anders deuten, als durch die Annahme, dass während einer bestimmten Epoche Bedingungen vorlagen, unter denen eine selektive Diffusion aus dem Schwerefeld stattfinden konnte. Die pro Zeitinterwall *dt* aus dem Schwerefeld eines

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[2] Herrn stud. B. Franck danke ich für einige ergänzende Versuche, die er in diesem Zusammenhang zur Bestimmung der Xenonadsorbtion an Quarzsand ausgeführt hat.

Planeten entweichende Menge-dN eines Gases ist eine Funktion der Masse M der Gasteilchen, die näherungsweise gegeben ist durch den Ausdruck (näheres z.B. bei J. H. Jeans, 1923, od. R. H. Fowler, 1929):

$$-\,dN = N_t \cdot A \cdot e^{-Mgr/RT}. \qquad (1)$$

Hierin bedeutet: N eine Anzahl von Molekeln, N_t die zur Zeit t vorhandene Anzahl und M das Molgewicht des betrachteten Gases, g die Schwerebeschleunigung in der Entweichungszone und r deren Abstand vom Gravitationszentrum, ferner R die Gaskonstante, T die absolute Temperatur und A eine im Wesentlichen von M unabhängige Konstante. Für die nach der Zeit t noch im Schwerefeld des Planeten vorhandene

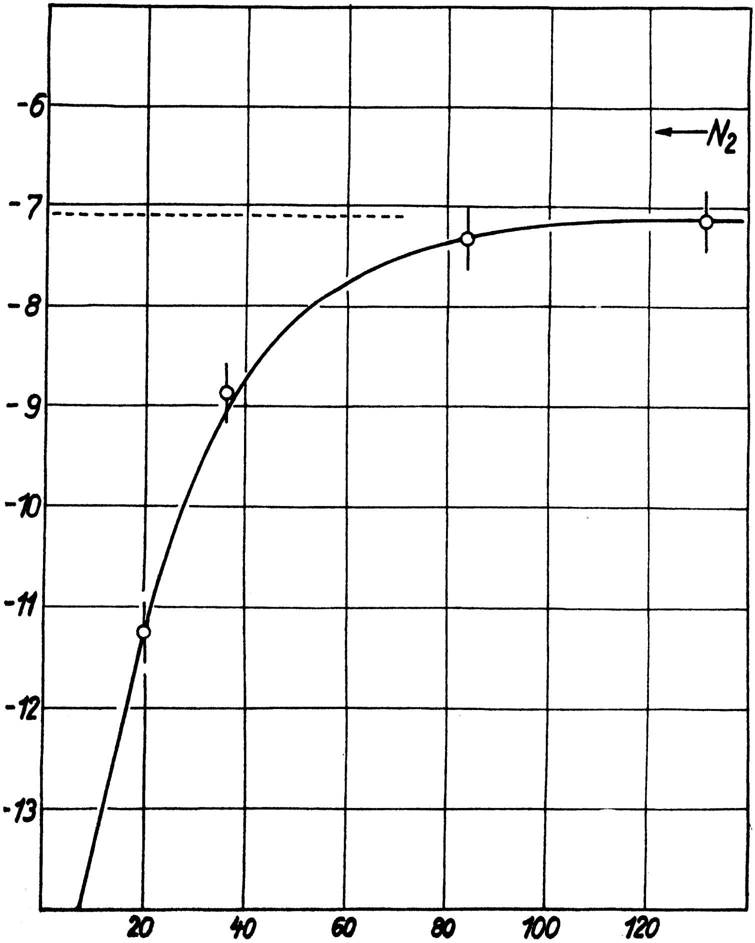

Abb. 1.—*Abszisse:* Atomgewicht M. *Ordinate:* Dekadischer Logaritmus des Verhältnis der Häufigkeiten der Edelgase in der Erdatmosphäre und im Kosmos ($\log N_{ter}/N_{sol}$), bezogen auf die Häufigkeit des Siliziums. Die ausgezogene Kurve entspricht der Funktion:

$$-\log N_{ter}/N_{sol} = 10 \cdot e^{-0{,}045\, M/m_1} + 7{,}1\,.$$

Menge eines Gases mit dem Molekulargewicht M gilt somit:[3]

$$-\ln N_t = t \cdot A \cdot e^{-Mgr/RT} - \ln N_0 \,. \quad (2)$$

Die gegenwärtige Häufigkeitsverteilung der Edelgase in der Atmosphäre entspricht nun in der Tat einer solchen Funktion, und zwar entspricht die in Abb. 1 eingezeichnete Kurve der Beziehung:[4]

$$-\ln \frac{N_{\mathrm{ter}}}{N_{\mathrm{sol}}} = 23 \cdot e^{-0,045 M/m_1} + 16.3 \,, \quad (3)$$

oder in dekadischen Logarithmen:

$$-\log \frac{N_{\mathrm{ter}}}{N_{\mathrm{sol}}} = 10 \cdot e^{-0,045 M/m_1} + 7.1 \,. \quad (4)$$

Für $N_t = N_{\mathrm{ter}}$ ergibt ein Vergleich von Gl. (2) mit (3) bezw. (4) die folgenden Werte:

$$N_0 = 10^{-7,1} \cdot N_{\mathrm{sol}} \,,$$

$$\frac{Mgr}{RT} = 0,045 \frac{M}{m_1} \,,$$

$$t \cdot A = 23 \,.$$

Den genauen Zahlenwerten soll wegen der vielen Unsicherheiten kein besonderes Gewicht beigelegt, sondern es sollen nur die Grössenordnungen diskutiert werden.

Vor allem ist der ausserordentlich kleine Wert des Exponenten Mgr/RT von Interesse. Für den gegenwärtigen Zustand der Erde erhält man mit einer Atmosphärenhöhe von rund 500 km (vergl. R. H. Fowler, 1929, S. 396) und einer Temperatur von rund 300° K für den Exponenten $Mgr/RT \approx 30\ M/m_1$. Merkliche Bruchteile schwererer Gase ($M/m_1 > 10$) können unter diesen Bedingungen in den Zeiträumen seit der Bildung der Erde nicht entwichen sein. Bei höherer Temperatur ist der Wert des Exponenten kleiner, jedoch wird selbst bei einer Temperatur von 3000° K und einer dieser Temperatur entsprechenden Atmosphärenhöhe von rund 15 000 km der Exponent $Mgr/RT \sim 1\ M/m_1$ (Näheres bei Russell u. Menzel, 1933). Ein hoher Atmosphärendruck vergrössert, wie man leicht sieht, die Atmosphärenhöhe nicht massgeblich. Dass sich das Entweichen der Edelgase bei noch höherer Temperatur vollzogen hat, ist nicht möglich, da dann die Silikate und andere Substanzen der Erdoberfläche bereits in hohem Masse flüchtig sind.

Es bleibt somit nur die Möglichkeit, dass das Gravitationspotential bezw. die effektive Schwerebeschleunigung während einer gewissen Zeitdauer kleiner war, als heute. Eine derartige Annahme scheint zunächst für die "Planetesimalhypothese" zu sprechen, nach der sich die Erde aus kleineren Körpern gebildet haben soll. In der Tat kann die Atmosphäre eines Himmelskörpers von etwa 1 bis 0, 1 % der Erdmasse unter geeigneten Bedingungen einen Zustand durchlaufen, in dem die Häufigkeitsverteilung der Edelgase ihrem gegenwärtigen irdischen Häufigkeitsverhältnis entsprechen würde. Dieser Zustand wird jedoch, wie eine Abschätzung der Geschwindigkeit der Abdiffusion (Näheres z.B. bei E. A. Milne, 1923) zeigt, bereits nach so kurzer Zeit durchlaufen, dass man diesen Körpern vom Zeitpunkt ihrer Bildung aus Solarmaterie bis zu ihrer Vereinigung zum Erdkörper nur eine Lebensdauer von wenigen Stunden oder allenfalls Tagen zubilligen könnte.

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[3] Derselbe Funktionsverlauf gilt auch für den Fall vollständiger Ionisation der Gasschicht, aus der das Entweichen stattfindet (Näheres hierüber bei W. Kuhn, 1948). Bei vollständiger einfacher Ionisation tritt in Gl. (1) und (2) anstelle von M im Exponenten die Grösse $\overline{M} - \overline{m}/2$, wobei $\overline{m}$ das mittlere Molgewicht der die Atmosphäre bildenden Gase bedeutet. In diesem Fall erfolgt somit das Entweichen lediglich um den für alle in der Atmosphäre enthaltenen Gase gleichen Faktor exp ($\overline{m}gr/RT$) rascher.

[4] Aus dimensionellen Gründen ist hier das Atomgewicht M durch die Atomgewichtseinheit m_1 dividiert.

Diese Schwierigkeit sowie die bekannten allgemeinen Einwände gegen die "Planetesimaltheorie" lassen es plausibler erscheinen, zur Deutung des kleinen Wertes des Exponenten die hohe Rotationsgeschwindigkeit der Erde zur Zeit ihrer Bildung in Betracht zu ziehen. Wie H. Jeffreys (1924) gezeigt hat, verlangsamt die Erde ständig unter dem Einfluss der durch den Mond verursachten Gezeitenreibung ihre Rotation. Sie hat sich in der ersten Zeit nach ihrer Entstehung sicherlich in weniger als 6 Stunden um ihre Achse gedreht. Für eine wie ein starrer Körper rotierende Atmosphäre verschwindet bei dieser Rotationsgeschwindigkeit die effektive Schwerebeschleunigung am Äquator bereits in einer Höhe von rund 10 000 km. Man kann leicht abschätzen, dass eine Atmosphäre, deren Gasmenge der gegenwärtigen entspricht, bereits bei Temperaturen von etwa 2000° K in diese Gegend hinaufreicht.

Es sei deshalb das folgende Bild vom Vorgang der Edelgasabwanderung vorgeschlagen, das sich vermutlich am ehesten in ein umfassenderes Bild der Planetenentstehung einordnen dürfte:[5] In einer ersten Phase nach der Kondensation der Erdmaterie (vergl. A. Eucken, 1944) ist der Hauptteil der in der ursprünglichen Solarmaterie enthalten gewesenen Edelgase, und zwar das rund 10^7-fache des zurückbleibenden Restes infolge der hohen Rotationsgeschwindigkeit, vermutlich auch zum Teil noch als dynamischer Bestandteil einer das ganze Planetensystem umfassenden turbulenten Gasmasse (im Sinne v. Weizsäckers, 1944), konvektiv, ohne das Auftreten von Separationseffekten von dem kondensierten Anteil der Erdmaterie abgetrennt worden. In einer sich anschliessenden zweiten Phase wurde bei abnehmender Atmosphärenhöhe und Rotationsgeschwindigkeit ein Stadium erreicht, in dem die Gasatome das Schwerefeld der Erde am Äquator nur mehr über einen, wenn auch kleinen Potentialwall verlassen konnten. Die zu Beginn dieses Stadiums vorhanden gewesene Neonmenge würde unter heutigen Bedingungen einem Partialdruck von etwa 200 mm Hg entsprechen. Der Gesamtvorgang würde einer allgemeinen Eigenschaft rotierender kosmischer Gasmassen entsprechen, die nach v. Weizsäcker (1948) sich stets in einen Kern und einen ins Unendliche entweichenden Teil auflösen.

In dem Stadium des Überganges der Abwanderung der Edelgase durch Konvektion in Abdiffusion kann sich bei abnehmender Temperatur und Atmosphärenhöhe die Separation der Edelgase vollzogen haben. Es ist kaum möglich, quantitativere Aussagen über den Verlauf des Prozesses zu gewinnen, da alle massgebenden Parameter, wie Erdrotation, Temperatur und Atmosphärenmenge, sowie ihre zeitlichen Änderungen in weiten Grenzen unbestimmt sind. Es scheint jedoch unmittelbar plausibel, dass ein derartiger Prozess näherungsweise eine Verteilung nach Gl. (3) ergeben kann.

Abschliessend sei noch darauf hingewiesen, dass eine derartige nach dem Atomgewicht differenzierte Abdiffusion der Edelgase aus dem Schwerefeld der Erde zwangsläufig auch zu einer Verschiebung der isotopischen Zusammensetzung der zurückbleibenden Gase geführt haben müsste, da Gl. (3) natürlich auch für Isotope Atomarten gilt. Man wird daher vermuten, dass die Edelgase der Erdatmosphäre etwas reicher an ihren schwereren Isotopen sind, als im

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[5] Vergl. Zeitschr. f. Elektrochemie (1949), Bd. 53, S. 237.

Kosmos. Im Falle des Neons haftet dieser Überlegung freilich noch eine zusätzliche Unsicherheit an: Es wäre möglich, dass die Erdatmosphäre unmittelbar nach der Edelgasabwanderung noch ärmer an Neon war, als heute und einen Teil des Neon durch nachträgliche Aufnahme geringer Mengen von unentmischter Solarmaterie erhalten hat. In Abb. 1 würde dann der Punkt für den ursprünglichen Wert des Neon noch tiefer liegen und der gegenwärtige Wert würde sich nur zufällig in den glatten Funktionsverlauf einordnen. Nimmt man jedoch an, dass alles atmosphärische Neon am Entmischungsprozess beteiligt war, dann erhält man nach Gl. (4) die in Tab. 2 angegebene isotopische Zusammensetzung des kosmischen Neon (s. Tab. 2).

TABELLE 2

	Massenzahl		
	20 (%)	21 (%)	22 (%)
Atmospärisches Ne.........	90,51	0,28	9,21
Kosmisches Ne (berechn.)...	93,4	0,2	6,4

Für die anderen Edelgase gilt die für Neon gemachte Einschränkung jedoch nicht. Für das Verhältnis $A^{36}:A^{38}$ müsste sich noch ein deutlicher Unterschied zwischen atmosphärischem und kosmischem Argon feststellen lassen. In Tab. 3 sind die Daten für die isotopische Zusammensetzung von Argon angegeben. Die Werte der zweiten Zeile erhält man, wenn man von dem aus K^{40} gebildetem A^{40} absieht und nur den Gehalt an primär vorhanden gewesenem A^{40} berücksichtigt, der durch Interpolation aus den Daten für die kosmische Häufigkeit von S^{36} und Ca^{44} abgeschätzt werden kann (Suess, 1946). Die Werte der dritten Zeile sind wie oben nach Gl. (4) berechnet.

In der Literatur finden sich mehrfach Angaben über einen auffallend hohen Neongehalt von natürlichen Gasen und Gesteinen (A. Gautier, 1909; W. Ramsay, 1912; Lord Rayleigh, 1938). Es ist möglich, dass in manchen Tiefengesteinen noch Edelgasmengen enthalten sind, die nicht oder nur teilweise den Separationsprozess der zweiten Phase der Edelgasabwanderung durchgemacht haben und deswegen vorwiegend aus Neon bestehen. Die Kenntnis der isotopischen Zusammensetzung von Neon und Argon aus solchen natürlichen Gasen und Gesteinen könnte eine Entscheidung in diesen Fragen bringen. Es ist auch mit der Möglichkeit zu rechnen, dass die beobachteten Schwankungen in der isotopischen Zusammensetzung des Kohlenstoffs (Näheres bei K. Rankama, 1948) zum Teil durch die Trenneffekte bei der Abwanderung der flüchtigen Kohlenstoffverbindungen aus dem Erdschwerefeld hervorgerufen wurden. Hier sind jedoch die Verhältnisse durch die Lage der Isotopenaustauschgleichgewichte besonders verwickelt. Nach Drucklegung der vorliegenden Arbeit ist dem Verfasser der Artikel von Harrison Brown "Rare gases and the formation of the earth's

TABELLE 3

	Massenzahl		
	36 (%)	38 (%)	40 (%)
Atmosphärisches A....	0,306	0,061	99,6
Ds. ohne radiogenes A.....	83,0	16,5	∼ 0,5
Kosmisches A (berechn.)...	85,6	14,1	∼ 0,4

atmosphere"[6] bekannt geworden, dem gleichfalls ein Vergleich der kosmischen und terrestrischen Edelgashäufigkeiten zugrundegelegt ist. Die hier dargelegten, und die von Harrison Brown ausgeführten Überlegungen beruhen praktisch auf den selben Grundannahmen und ergänzen sich in vielerlei Hinsicht.

Den Herren P. Harteck und H. Jensen danke ich herzlich für wertvolle Diskussionen über dieses Thema. Es sei mir auch gestattet, der Geological Society of America an dieser Stelle zu danken für die Vermittlung privater Unterstützungen im vergangenen Jahre, durch die mir damals die Weiterführung wissenchaftlicher Arbeiten ermöglicht wurde.

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[6] Aus: "The atmospheres of the earth and planets" Ed. G. P. Kuiper, Chicago, University of Chicago Press, 1949, 260 pp.

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A NOTE ON THE GEOCHEMISTRY OF THALLIUM[1]

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ABSTRACT

The manner of occurrence of thallium in the upper lithosphere is discussed. The value 0.6 g/ton is tentatively adopted for the abundance of thallium in igneous rocks. The presence of notable variations in the Rb/Tl ratio in igneous rocks is established. The geochemical coherence of the pair Rb-Tl is discussed and compared with the coherence in the pairs Ga-Al, Cb-Ta, and Zr-Hf, and in the group of lanthanum and the lanthanides. It is concluded that zirconium and hafnium probably form a more coherent pair than do rubidium and thallium.

INTRODUCTION

So far, thallium has been and partly still remains one of the geochemically more imperfectly known elements. The most recent contributions to the geochemistry of thallium consist of a series of papers by Ahrens (1945*a*, *b*; 1947*a*, *b*; 1949*a*, *b*) which have shed much welcome light on the mineralogical geochemistry of this metal. Ahrens emphasized the peculiar chemical character of thallium: the unipositive thallium ion behaves chemically much like the ions of the alkali metals of higher atomic weight. He analyzed spectrochemically 167 mineral samples for their content of Rb_2O and Tl_2O. This material consisted of primary and hydrothermal pegmatitic microclines, amazonite, plagioclase, muscovite, phlogopite, lepidolite, zinnwaldite, biotite, pollucite, rhodizite, and cesian beryl. The results revealed a very close and unique association of rubidium and thallous ("alkali-metal") thallium in potassium minerals. The Rb_2O/Tl_2O ratio was found to lie between 12 and 650; the mean ratio was 100. Ahrens found no shift in the Rb/Tl ratio "throughout the selective crystallization of minerals." Therefore, he concluded, rubidium and thallous thallium comprise a pair of elements which are remarkable for the constancy of their relative proportions; the relationship apparently is independent of the mineral type; it holds over all ranges of concentration; moreover, if the analyses are confined to a single area or source, the Rb_2O/Tl_2O ratio remains almost exactly constant within the limits of experimental error and even approaches an isotopic constancy. Consequently, "alkali-metal" thallium and rubidium are very closely associated geochemically in the upper lithosphere, and it is possible that their relationship is closer than the relationship between any other pair of elements, with the possible exception of zirconium and hafnium and some of the rare-earth metals. Moreover, Ahrens used the newly established Rb_2O/Tl_2O ratio to calculate a new value for the abundance of thallium in the upper lithosphere because he thought that the close association of thallium and rubidium in minerals would make possible a reasonable estimate of the abundance of thallium, derived from the abundance of rubidium. For the abundance of rubidium in igneous rocks Ahrens used Goldschmidt's (1937*a*, *b*) value of 300 g/ton (actually, 310 g/ton), which he thought to be reasonably ac-

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curate; and for the abundance of thallium in igneous rocks he adopted the value 3 g/ton, which is ten times as high as the old value, or 0.3 g/ton, of Goldschmidt (1937*a*, *b*). It may be added here that the previous abundance values of thallium in the upper lithosphere calculated by Noddack and Noddack (1930, 1934) are still lower, viz., 0.1 and 0.01 g/ton, respectively. Ahrens claimed that Goldschmidt's value is based on relatively scant data and found that the old value was evidently too low. Finally, Ahrens pointed out that the new value refers to the "alkali-metal" thallium; but, because the content of thallium present in certain sulfide minerals is insignificant when compared with the thallium content of the silicate minerals of potassium, the new value may be regarded as applying to the total abundance of thallium in the upper lithosphere.

The purpose of this paper is to reconsider the newest abundance values presented for thallium and to compare the pair rubidium-thallium with some other pairs of elements closely associated in the upper lithosphere.

MANNER OF OCCURRENCE OF THALLIUM IN THE UPPER LITHOSPHERE

Geochemically, thallium found in the upper lithosphere is characterized by its property of becoming enriched in late crystallates. In this respect it behaves like the heavy alkali metals, rubidium and cesium. The most important manner of occurrence of thallium in the upper lithosphere is its camouflage in silicate minerals rich in potassium. Potash feldspar and the micas, muscovite and biotite, evidently are the main seats of thallium in the silicic calcalkalic igneous rocks and particularly in their pegmatites. According to Ottemann (1941), thallium is also rather strongly enriched in the minerals of pneumatolytic tin-bearing rocks. Its occurrence in potash feldspar is explained by the fact that thallous thallium is able to form a feldspar structure, as the thermochemical calculations of Schiebold (1931) show. In potash feldspar, Tl^+ (radius 1.49 kX), like Rb^+ (radius 1.49 kX), substitutes for K^+ (radius 1.33 kX). No independent minerals of Rb^+ and Tl^+ are known to exist. In plagioclase feldspars, Tl^{3+} (radius 1.05 kX) may become captured by Ca^{2+} (radius 1.06 kX), as Ottemann (1941, p. 166) has suggested; in fact, Ottemann (1941, p. 162) reports as much as 10 g/ton thallium in plagioclase, whereas the maximum content found by Ahrens (1948*b*) is 2.9 g/ton thallium. In potash-rich silicate minerals, particularly in potash feldspar, thallium also accompanies cesium and lead. In like manner a geochemical alliance with lead characterizes the occurrence of thallium in sulfide minerals. Evidently, the most important site of thallium among sulfides is galena, in which it replaces lead. A similar replacement exists also in some plumbiferous sulfosalts, such as sartorite, $PbS \cdot As_2S_3$, dufrenoysite, $2PbS \cdot As_2S_3$, and jordanite, $4PbS \cdot As_2S_3$. The thallium minerals proper: lorandite, crookesite, vrbaite, and hutchinsonite, which are sulfides, selenides, and sulfosalts, are very rare and consequently do not possess much geochemical significance. Oftedal (1940) found the common sulfide minerals, with the exception of galena, devoid of thallium. The commonly reported occurrence of thallium in pyrite is probably erroneous (V. M. Goldschmidt, according to Oftedal, 1940). The average thallium content of primary magmatic sulfides, according to the analyses of Noddack and Noddack (1931), is 1 g/ton thallium. It is evident that the content of thallium in sulfide minerals is too low to be of geochemical importance and to affect the abundance of thallium in the upper lithosphere.

CONTENT OF THALLIUM IN IGNEOUS ROCKS

Ahrens (1948*b*) failed to determine the content of thallium in granites and other rocks relatively rich in potassium because their thallium content was below the limit of detection of the spectrochemical method that he used. However, there is a number of direct previous determinations of the thallium content of igneous rocks, made by V. M. Goldschmidt and H. Hörmann (Goldschmidt, 1937*b*, p. 88), Preuss (1941, p. 19), and Ottemann (1941, p. 156). All these authors used spectrochemical methods of investigation; in order to increase the sensitivity, Preuss distilled the readily volatile elements from rock samples in a carbon-tube furnace directly into the carbon arc, and Ottemann heated the samples in a high-vacuum electric furnace, whereby sublimates were obtained for further studies. The results of their analyses are presented in table 1.

TABLE 1

CONTENT OF THALLIUM IN IGNEOUS ROCKS

Rock	Tl (g/ton)
Granites and liparites, average (Goldschmidt and Hörmann in Goldschmidt, 1937*b*)	0.5–1
Composite of 11 German gabbros (Preuss, 1941)	0.3
Composite of 14 German granites (Preuss, 1941)	3
Granite, Eibenstock, Harz, Germany (Ottemann, 1941)	1.0
Granite, Schellerhau, Harz, Germany (Ottemann, 1941)	1.0
Granite, Epprechtstein, Harz, Germany (Ottemann, 1941)	2.0
Granite, Brocken, Harz, Germany (Ottemann, 1941)	2.1

With due consideration to the scarcity of the analyses presented in table 1, the values show that the thallium content of subsilicic rocks is considerably lower than the content in silicic rocks. The high average reported by Preuss for the composite mixture of German granites may be attributed to the geochemical peculiarities of the Caledonian and Variscan granites analyzed. Some other trace elements are also known to be present in exceptional quantities in these granites. The average of Ottemann's analyses for the Harz granites is 1.5 g/ton thallium, a value which compares rather well with the average reported by Goldschmidt and Hörmann for granites and liparites. The values presented in table 1 clearly show that the new value given by Ahrens for the abundance of thallium in igneous rocks is much too high. But table 1 also shows that Goldschmidt's value is somewhat too low; the value 0.6 g/ton might be chosen to represent the true abundance of thallium in igneous rocks at least tentatively until more analyses become available.

The abundance of thallium and its neighbors in the Periodic System is given in table 2.

According to the rule of Oddo and Harkins, both mercury and lead should be more abundant than thallium. The abundance relationships of lead (abundance according to Goldschmidt, 1937*b*) and thallium certainly follow the rule, whereas in the case of mercury and thallium the relationship seems to be more obscure. The value 0.5 g/ton for mercury is based on determinations from sedimentary rocks (Goldschmidt, 1937*b*), whereas the lower value is the recent estimate of Saukov (1946); this value agrees well with the analyses of igneous rocks made by Stock and Cucuel (1934), Preuss (1941), and Saukov himself. The discrepancy may be explained by the assumption that mercury is geochemically comparable with the atmophile elements which were largely lost during the early stages of the evolution of the earth. Therefore, the present terrestrial abundance of mercury is abnormally low. In like manner mercury is considerably more deficient in the meteorites than the rules for the cosmic abundance of its isotopes require (Suess, 1947, p. 604). As Goldschmidt (1937*b*, pp. 11, 84) points out, much of the mercury in the sediments and their derivatives probably has been produced by the degassing of the upper lithosphere, i.e., by volcanic emanations and hydrothermal solutions. If Ahrens' new value for thallium is considered, the deviation from the rule of Oddo and Harkins grows still more pronounced than with the lower values, and therefore it follows that a lower value is to be preferred. The rule of Oddo and Harkins, which is no selection principle but a true natural law, affords a welcome check of the abundance of the elements considered.

TABLE 2

Z	Element	Abundance in the Upper Lithosphere (g/ton)
80	Hg	0.077–0.5
81	Tl	0.6
82	Pb	16

THE Rb/Tl RATIO IN IGNEOUS ROCKS

According to Ahrens, the Rb/Tl ratio does not change during the crystallization of minerals. In order to study the Rb/Tl ratio in some igneous rocks, table 3 is presented; it is compiled from analyses available in literature. The first two rubidium values are from Goldschmidt, Berman, Hauptmann, and Peters (1933, p. 238); the third value is a redetermination by Goldschmidt, Bauer, and Witte (1934, p. 51). The thallium values are those listed in table 1, and the Rb/Tl ratio for the potassium-rich minerals of granite pegmatites is calculated from the mean Rb_2O/Tl_2O ratio given by Ahrens (1948*b*, p. 580).

Table 3 shows that considerable variation in the Rb/Tl ratio occurs during the magmatic differentiation. Certainly, no conclusion on the abundance of thallium can be based on the assumption of the prevalence of a stable Rb/Tl ratio in igneous rocks and their minerals, and therefore the evidence of table 3 also contradicts the new high abundance value for thallium.

TABLE 3

CONTENT OF RUBIDIUM AND THALLIUM AND THE Rb/Tl RATIO IN SOME IGNEOUS ROCKS

Rock	Rb (g/ton)	Tl (g/ton)	Rb/Tl (Weight Ratio)
Composite of 11 German gabbros	18	0.3	60
Composite of 14 German granites	550	3	183
Composite of 14 German granites	830	3	277
Granite pegmatites (potassium-rich minerals)	......	......	95

The discrepancy between Ahrens' high value and the other, lower values is readily explained by the fact that the Rb/Tl ratio adopted by Ahrens as the universal ratio actually is valid only for the potassium-rich minerals of late crystallates, in particular of granite pegmatites, and not for the minerals of the igneous and quasi-igneous sequence as a whole. Moreover, the abundance of rubidium is not known with certainty, and therefore an abundance value for thallium calculated from the Rb/Tl ratio remains necessarily uncertain. It must be emphasized that the abundance values given by Goldschmidt (1937*a*, *b*) for a number of elements are based on analyses of Caledonian and Variscan igneous rocks and consequently may be uncertain because there is much reason to believe that regional differences in the content of many trace elements exist in the upper lithosphere.

THE COHERENT PAIR Rb-Tl COMPARED WITH SOME OTHER PAIRS OF ELEMENTS

Ahrens has emphasized that, with the possible exception of the pair Zr-Hf and, perhaps, some of the rare-earth metals, no other pairs of elements are known which are so closely associated as are "alkali-metal" thallium and rubidium in silicate minerals. With this statement Ahrens evidently refers to the concept of geochemical coherence: two or more elements are geochemically coherent if they constantly follow each other in nature, not only in igneous rocks but in the endogenic and exogenic cycle of terrestrial matter in general. A convenient measure of the geochemical coherence is the variation of the abundance ratio of the elements concerned in different materials. In the following the degree of coherence in some pairs of elements will be considered.

Gallium-aluminum.—As Goldschmidt and Peters (1931) showed, gallium always follows aluminum in nature. They found that in minerals the atomic Ga/Al ratio varies from about 1 : 10 to 1 : 120,000—or by a factor higher than 10^4. Consequently, the degree of coherence in the pair Ga-Al is low.

Lanthanum and the lanthanides.—This group of elements was formerly considered to be highly coherent, but recent research has proved that notable differences exist between the distribution of the various lanthanides, which may in certain cases become separated from each other (for a recent summary see Haberlandt, 1948).

Columbium-tantalum.—It is now known that columbium and tantalum actually may become quantitatively separated from each other in minerals (Rankama, 1944, 1948). Therefore, these metals do not form a fully coherent pair, even though they usually follow each other in nature.

Zirconium-hafnium.—This pair certainly is highly coherent: where zirconium does not occur, there hafnium is absent as well. Zirconium, consequently, is a very powerful protective element (*Schutzelement*) for hafnium. No hafnium is present in titanium and thorium minerals. It is an exclusive constituent of zirconium minerals. According to Hevesy (1925), the HfO_2/ZrO_2 ratio varies from 0.007 (in fava) to 0.5 (in thortveitite). The corresponding atomic Hf/Zr ratios are 0.004 and 0.3, respectively, and therefore the ratio varies by a factor of 75. The coherence is strong.

Rubidium-thallium.—According to Ahrens (1948*b*), the Rb_2O/Tl_2O ratio varies from 11.3 (in muscovite No. 165) to 660 (in potash feldspar No. 16). The corresponding atomic Rb/Tl ratios are 27.3 and 1434.7, respectively, and the Rb/Tl ratio varies by a factor of about 53. It is evident that the coherence is strong.

A comparison of the data presented above shows that the coherence in the pairs Zr-Hf and Rb-Tl is higher than in the pairs Ga-Al and Cb-Ta, and probably also higher than in the lanthanide group, even though here the lack of analytical data so far does not allow a direct comparison. On the other hand, the coherence in the pairs Zr-Hf and Rb-Tl is of the same degree of magnitude. Even though the geochemistry of hafnium is known very inadequately, all information available shows that hafnium never parts with zirconium in nature.

Of course, the values presented above allow a comparison of coherence only with reference to Zr-Hf– and Rb-Tl–bearing minerals of late crystallates, but it is probable that rubidium and thallium actually are a less coherent pair than zirconium and hafnium. The reason is that, all manners of occurrence and all valence states of thallium being considered, a complete separation of rubidium and thallium takes place in the thalliferous sulfide minerals. However, if only univalent thallium is considered, one may state, in material agreement with Ahrens, that the association of thallium and rubidium is, indeed, remarkably close, in the potash-rich silicate minerals of granite pegmatites and of silicic igneous rocks.

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GEOLOGICAL NOTES

TRIVOLI SANDSTONE OF WILLIAMSON COUNTY, ILLINOIS[1]

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New information on the nature of the sandstones of the lower part of the McLeansboro group of the Pennsylvanian system in a part of southern Illinois has resulted from recent diamond drilling in northern Williamson County, in the southern Illinois coal field (fig. 1). A sandstone which crops out in the immediate vicinity of the bore holes is the uppermost sandstone encountered in the cores, and a definite correlation can be made from surface exposures to the subsurface section. Two sets of samples of this sandstone were collected, one from the drill holes, the other from the outcrop. Thin sections were cut and heavy mineral analyses made from these samples.

The sandstone sampled is the basal member of the cyclothem containing No. 8 coal. It lies unconformably on the uppermost shale member of the cyclothem next below. Its base is about 200 feet above the base of the McLeansboro group. Kosanke and Eddings (1946, p. 1212) have correlated coal No. 8 of southern Illinois with the Trivoli (No. 8) coal of western Illinois. Thus the basal sandstone member of the Williamson County cyclothem presumably can be correlated with the basal member of the Trivoli cyclothem, the Trivoli sandstone. Cooper (1945, p. 16) has correlated the Trivoli sandstone of western Illinois with the basal member of the Missouri series of the Mid-Continent section, lying just above a major floral and faunal break in the Pennsylvanian strata.

On the outcrop the sandstone is brown to reddish-brown, uniformly fine-grained, weathering into large slabs and flags, and uniform in appearance from top to bottom. The sandstone is prominently marked with oscillation ripples; the amplitude averages 2 cm., and the wave-length ranges from 8 to 12 cm. Cross-bedding of the aqueous type is common. The angle of the foreset beds is about 10°–15°. Characteristically the rock is case-hardened by hematite and limonite as much as 2 cm. from the surface. The basal conglomerate commonly present in the cores was not exposed in any outcrop.

In the core the sandstone is light gray, with a weathered brown zone at the top, commonly 2–5 feet thick. Shaly micaceous carbonaceous partings are not uncommon, and ripples are prominent. The top of the sandstone grades into the overlying silty shale member of the cyclothem through a transitional zone, 2–5 feet thick, characterized by thin shaly laminae in the sandstone, which increase in number and thickness upward. In all the drill holes the contact with the shale below is abrupt and uneven and, in many places, is inclined as much as 20°. The interval between the base of the sandstone and the top of the first coal bed below shows appreciable variation; the interval is commonly 50 feet, but in some places it decreases to 10 feet. It is probable that the Trivoli sandstone is unconformable on the shale below. In general, the sandstone is uniform throughout.

The sandstone ranges in thickness from 60 to 100 feet in northern Williamson County; in Franklin County, to the north, it reaches a maximum thickness of 157 feet. Approximately 6 miles north of the outcrop

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[1] Published with the permission of the Chief, Illinois State Geological Survey. Manuscript received February 18, 1949.

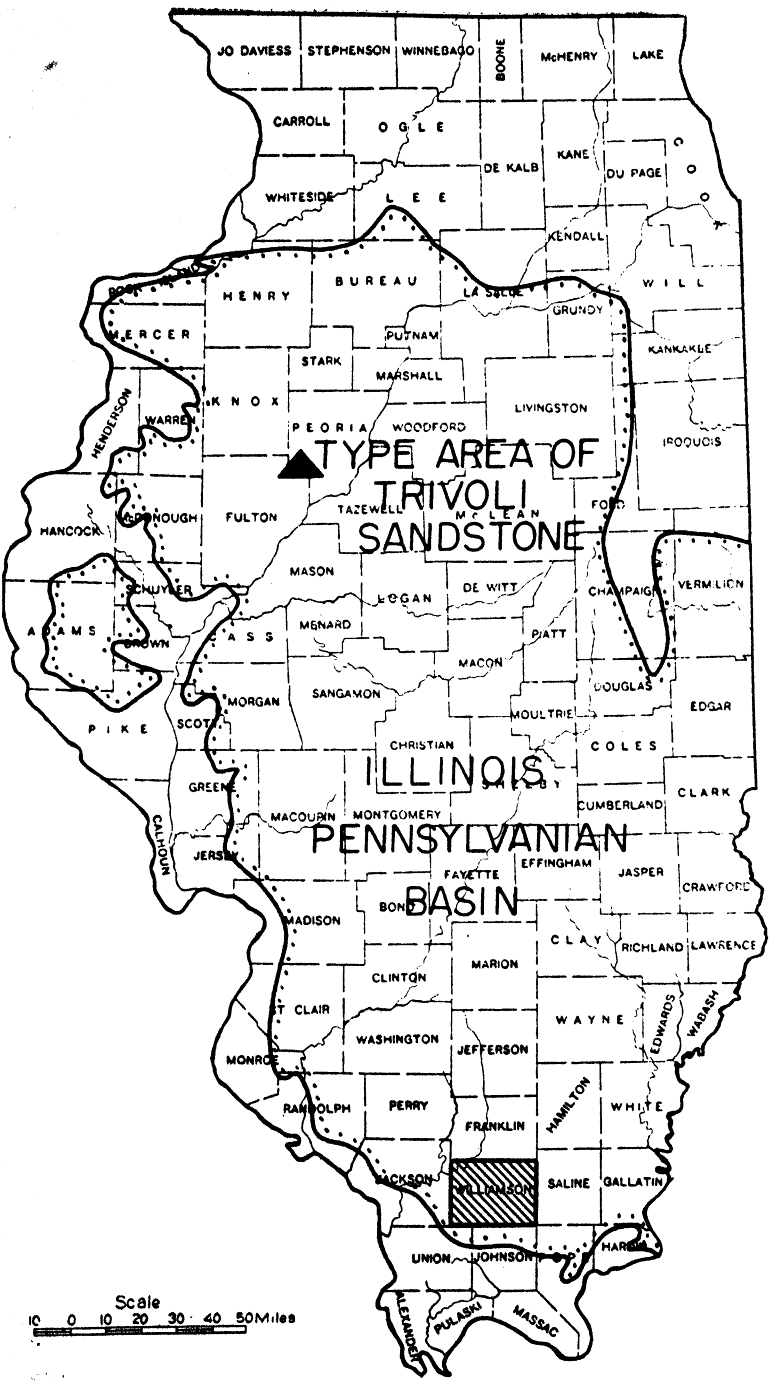

FIG. 1.—Index map, showing type area of Trivoli sandstone and location of Williamson County in Illinois.

area in Williamson County the sandstone grades laterally northward into sandy shale, and thence into shale. The sandstone is lost as an identifiable unit until it reappears some 60 miles to the north.

As shown in thin sections (pl. 1, *A*, *B*), the rock contains, on the average, 60 per cent detrital quartz grains. Next in order of abundance is the clay matrix, composed of sericite, chlorite, illite, and other clay minerals. The remainder of the rock consists of various types of feldspar, rock fragments, and chert. With a few exceptions, discussed below, chemical cements are less than 7 per cent. They include authigenic quartz, calcite, and siderite. Heavy mineral accessories are few and slightly varied, the most abundant being tourmaline and zircon.

The sandstone contains several types of detrital quartz. The most prominent is the normal igneous quartz which is free of strain shadows, has even edges, and contains small dusty inclusions. From 40 to 60 per cent of the detrital quartz is of igneous origin. Almost equally prominent are the metamorphic quartz grains which are lath-shaped, lensoid, and exhibit mild to pronounced strain shadows. Many of them are metamorphic quartzite fragments which show extreme cases of suturing of secondary quartz. Authigenic quartz as enlargement of detrital grains rarely exceeds 5 per cent of the total quartz. Second-cycle grains are scarce. The visual roundness averages about 0.25; metamorphic quartz is a little lower, and igneous quartz is a little higher, in roundness.

Feldspars, which constitute from 2 to 10 per cent of the rock, include sodic plagioclase, orthoclase, microcline, and a few grains of anorthoclase. The feldspars are slightly more rounded than the quartz and have a visual roundness of 0.5. Most of the feldspars are fresh, but some grains show extensive kaolinization.

Rock fragments, composing up to 15 per cent of the sandstone, are of many types of argillaceous sedimentary and low-rank metamorphic rocks and some chert. The most common fragments are phyllites. Many of the shale fragments are similar to Pennsylvanian shales lower in the section and thus may be of local origin.

The clay matrix constitutes about 20 per cent of the sandstone. It is exceedingly fine-grained, acting as a binder for the larger detrital grains. The most abundant mineral is sericite and, in order of decreasing abundance, chlorite, illite, and kaolinite. A few grains of biotite are present. The clay particles show no preferred orientation.

Muscovite is very prominent in the Trivoli sandstone. It occurs as long shreds and plates, many of them bent and broken, and shows conspicuous parallel orientation. Some samples contain as much as 5 per cent of muscovite.

Two carbonate minerals—calcite and siderite—occur as cement. The carbonate cement composes about 5 per cent of the rock, except for a few samples in which it is as much as 20 per cent. The carbonate occurs as irregular patchy areas and replaces quartz and feldspar. In the few samples with high carbonate content the rock has a relatively clean appearance; the clay matrix is at a minimum, and calcite is the dominant cement. This calcite has extensively replaced quartz and feldspar and is itself being replaced by authigenic quartz. Presumably the clean-washed permeable sand was more favorable to carbonate infiltration and precipitation. Sharply euhedral secondary quartz replaces carbonate. Siderite occurs as small irregular clusters of rhombohedrons

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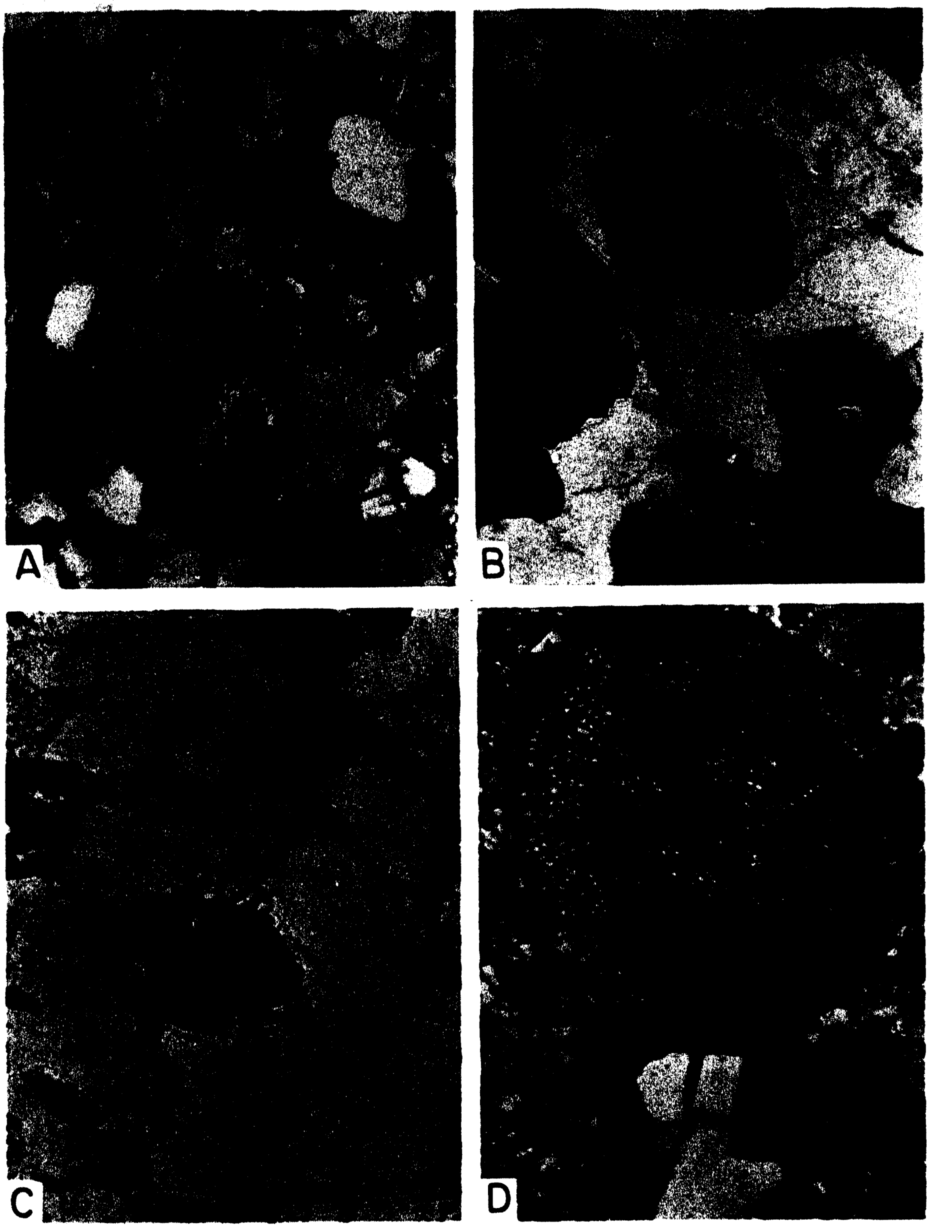
PLATE 1

A, Typical appearance of Trivoli graywacke. Quartz, plagioclase feldspar, and clay matrix are readily distinguishable. Crossed nicols. Magnification 100×.

B, "Washed" graywacke with calcite cement. Euhedral secondary enlargement of quartz replacing calcite cement. Crossed nicols. Magnification 100×.

C, Glauconite (*arrow*) in the Trivoli sandstone. Plane-polarized light. Magnification 100×.

D, Cross section of fusainized stem in the Trivoli conglomerate. Crossed nicols. Magnification 100×.



Thin sections of Trivoli sandstone

and as extensive replacement of calcite. The siderite is about 30 per cent of the carbonate cement. It is extensively altered in the outcrop to hematite and limonite.

A very small amount of glauconite (pl. 1, *C*) is present. As it shows no evidence of being worn in transport, it is considered to be authigenic. It occurs in samples with a greater amount of biotite than customary and, as suggested by Galliher (1935, pp. 1351–1366), may be the product of marine alteration of the biotite.

The basal conglomerate of the sandstone, found in about 65 per cent of the drill holes, is generally 2–3 feet thick, but rarely attains a thickness of 6 feet. The matrix and fine detrital material are much the same as in the nonconglomeratic sandstone, but the conglomerate has slightly more carbonate cement, as much as 15 per cent. The pebbles are irregular in shape, commonly flattened, with the long dimension ranging from $\frac{1}{4}$ inch to 2 inches. Almost all the pebbles are of clay ironstone concretions, such as are commonly found in shales above a coal bed in Illinois. Other pebbles are of a sandstone tightly cemented with hematite and limonite, an occasional limestone pebble, and rarely a coal pebble.

Noteworthy are the plant remains found in the conglomerates (pl. 1, *D*). They are not impressions but are large fragments of stems and twigs whose abraded edges show evidence of transport. Invariably, the material is fusain or semi-fusain. The intracellular pore spaces of the fusain are open, in contrast to the fusain in Illinois coal beds, which is often mineralized with pyrite.

The mineralogy of the Trivoli sandstone is in sharp contrast to the mineralogy of sandstones lower in the Pennsylvanian section in Williamson County, although the Trivoli sandstone of Williamson County is similar to the sandstones between the Isabel sandstone at the base of the Carbondale group and the Trivoli sandstone of western Illinois. The lower sandstones in Williamson County average 80 per cent detrital quartz and 20 per cent chemical cement; in general, they are washed clean. They are typical representatives of the clean quartz sands of the orthoquartzite series. Thus the mineralogy of the Trivoli sandstone may be a reflection of the uplift and renewed erosion of the source area. The conglomerates of the lower Pennsylvanian in southern Illinois are very different from the Trivoli conglomerate. The former consist of thick massive beds containing well-worn quartz pebbles transported from a distant source; the latter is a thin basal conglomerate of slightly worn material of local derivation.

According to the cyclothem hypothesis of Pennsylvanian sedimentation, most of the basal sandstones of cyclothems have been considered to be continental deposits (Weller, 1930, p. 119). The few exceptions pointed out have been interpreted as "the shoreward facies of marine deposits contemporaneous with the aggradation of the alluvial plain" (Weller, 1930, p. 120). A major objection to the marine origin of most of the sandstones is the occurrence of autochthonous coal beds in them. No coal beds have been observed in the Trivoli sandstone of Williamson County. Petrographic and field evidence suggests that this sandstone in this particular area is more probably marine than nonmarine. The occurrence of wave ripple-marks, the uniformity of lithology, the thin basal conglomerate, and the glauconite are indications of a marine environment. This sandstone appears to constitute a marine facies of a dominantly continental deposit.

The sandstone is in the category of a low-rank graywacke or a subgraywacke. It implies contemporaneous moderate deformation in the source area of the sediments. The major source area has been considered to be "Appalachia," with minor amounts of detritus coming from Llanoria and the Wisconsin highlands. Investigations are in progress to establish more definitely the sources of sediment, the influence of tectonic movements of the source areas on sedimentation, and the amounts contributed by each source area to the sediments of the slowly downwarping Eastern Interior basin in Pennsylvanian time.

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A PROPOSED METHOD FOR THE MEASUREMENT OF GEOLOGIC TEMPERATURES[1]

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The establishment of an absolute geologic thermometer is of utmost importance in modern petrology. At present it is possible only in rare cases to determine with any degree of certainty the temperature at which a rock has been formed. If a means of temperature determination accurate to 100° C. and usable from 100° to 1000° C. were available, many of the problems of metamorphic and igneous petrology could be solved. For example, the question of the solution or replacement origin of pegmatite bodies might be answered if their temperature of emplacement, as well as the temperature of emplacement of the enclosing rocks, could be determined. One of the biggest problems of petrology, namely, the origin of granites, also hinges upon questions of temperature of formation. At present most methods of temperature determination are based upon the temperature stability of various minerals or mineral assemblages. These methods have the disadvantages of giving only a relative measure at best, and one the value of which is a function of composition as well as of temperature and pressure.

The following method has the advantage of being a first-order function of the temperature and pressure and depends on bulk rock composition in but a minor way.

The method, simply stated, is to measure the distribution ratio of a suitable minor element between two minerals in equilibrium in a rock. The distribution ratio should be determined solely by the temperature and pressure at which the distribution was established.

The minor element should be one that is present in large enough quantity for accurate and simple determination. This element should not be one that substitutes easily in one of the mineral lattices and not in the other, although this qualification may prove to be unimportant. One important requirement is that the minor element be present in such small amounts that the laws of the perfect solution can be applied.

H. C. Urey[2] has suggested the use of an ion which will substitute for an ion present in both minerals, i.e., germanium substituting for silicon in quartz and feldspar. Such a system is more likely to behave like a perfect solution than is the case in which large lattice distortions result upon substitution.

Because only small amounts of each phase can be obtained in a pure condition because of the fineness of the grain of many of the rocks in question and because the element to be determined is present in a concentration of 0.01–100 parts per million, a very accurate and sensitive analytical tool is necessary. The radiochemical method (Brown and Goldberg, 1949, p. 347) fits the above requirement. The work can be divided into several parts as follows:

1. An element must be found that exists

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[1] Manuscript received August 8, 1949.

[2] Personal communication.

in natural rocks in the proper concentration and whose distribution coefficient is not too far removed from unity. This element must be one whose concentration can be conveniently determined with sufficient accuracy. In addition, the slope of the temperature versus distribution coefficient curve must be sufficiently large that the proper accuracy can be obtained.

2. The temperature coefficient of the distribution ratio between two minerals—say, quartz and potash feldspar—must be determined. This may be done in two ways: (*a*) by the simultaneous synthesis of the two minerals in a closed system at a known temperature and pressure and (*b*) by equilibrating a natural rock at different temperature and pressure in the presence of water vapor.

3. Field investigations—a series of rocks from whose occurrence a relative temperature scale can be derived—must be investigated to see whether equilibrium relationships occur as far as the distribution coefficient is concerned.

4. The influence of solid solution and order-disorder changes in one host lattice on the distribution coefficient must also be investigated.

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REVIEWS

Cretaceous and Tertiary Subsurface Geology: The Stratigraphy, Paleontology, and Sedimentology of Three Deep Test Wells on the Eastern Shore of Maryland. (State of Maryland, Department of Geology, Mines, and Water Resources, Bull. 2 [1948].) Pp. 456; figs. 30; pls. 39. $2.00.

This bulletin is a comprehensive report on three deep tests drilled during the period 1943–1946 by the Ohio Oil Company, the Socony-Vacuum Oil Company, and the Standard Oil Company of New Jersey. Cuttings, cores, and electric logs were donated to the Department of Geology, Mines, and Water Resources, under the directorship of J. T. Singewald, Jr. Several collaborators conducted an exhaustive study of the data furnished by the wells, especially a complete core from a depth of 1,000 feet to bottom (5,568 feet), in the Ohio Oil Company—L. G. Hammond No. 1 well in Wicomico County, Maryland.

The first one hundred and twelve pages of the volume are a report on the subsurface Cretaceous and Tertiary geology by J. L. Anderson, based largely upon the cored well and tied in with other data. Detailed heavy-mineral and size analyses were run on core samples, affording a wealth of data in the form of histograms, statistical parameters of size, and graphs of the relations among the parameters. These data, in conjunction with the paleontological evidence, were used to reconstruct the environment of deposition of the sediments and to correlate the section among the three wells. Ditch samples from the Maryland Esso well were logged by R. M. Overbeck, and an appendix at the end of the volume presents the log data.

The second part of the volume describes the paleontology of the section penetrated by the cored well. Experts in several fields were engaged in the study, including Julia A. Gardner, Ann Dorsey (Mrs. F. W. Clapp), L. W. Stephenson, H. E. Vokes, K. E. Lohman, F. M. Swain, and J. A. Cushman, under arrangements made with Dr. J. B. Reeside, Jr., of the United States Geological Survey. The papers on paleontology include studies of mollusks, diatoms, ostracods, and Foraminifera. Many new species were found, which are described and illustrated in the thirty-nine accompanying plates. These papers provide an important reference on the paleontology of the East Coast Tertiary and Cretaceous, which will prove valuable also to workers along the Gulf Coast.

The data from the three wells, tied in with outcrop and other well information, show that the section thickens markedly from west to east. In Kent County, just east of Chesapeake Bay, the Tertiary is about 100 feet thick. In Worcester County, along the coast some 80 miles southeast, the Tertiary is 2,000 feet thick. The Cretaceous is 1,200 feet thick in Kent County, and nearly 6,000 feet thick in Worcester County. Between the Cretaceous and the Basement Complex east of Chesapeake Bay is a buried occurrence of Triassic rocks, consisting mainly of reddish-brown and apple-green shales, sandy shales, and arkosic sandstones. The correlation is based on lithologic similarity to the known Triassic of Marlyand, inasmuch as no fossils were found. The Triassic is 585 feet thick in Worcester County, where it is more shaly than in the Ohio Oil Company well farther southwest.

The conditions of sedimentation were reconstructed mainly from data afforded by the cored well. It was concluded that the beds in this well were deposited in a continental environment which changed gradually to deltaic and finally to marine. Marine conditions set in at the base of the Matawan formation (top of Upper Cretaceous) and continued uninterruptedly into the Tertiary. Progressive overlap of the beds toward the west is more pronounced in the Lower Cretaceous than in the upper part of the well section. Some brackish-water conditions in the Upper Cretaceous were noted in the easternmost of the three deep wells.

The petroleum potentialities of eastern Maryland were assessed from the well data, and it was concluded that, although the section is suitable for the accumulation of petroleum, it is not very favorable for petroleum generation.

If oil occurs in the Cretaceous, it is believed that it will be found east of the coast line, in more marine parts of the section. The Tertiary, although marine, has only shallow cover and seems unsuitable for oil occurrence. No evidence of petroleum was found in any of the sands in the three wells.

It is apparent from the detailed studies made upon the material and from the critical analysis of the data afforded that this volume provides a definitive treatment of the subsurface Tertiary and Cretaceous of the central part of the Atlantic Coast. Its value lies not alone in the insight that it affords into the geological history of an important area but also in the comparative data which it affords for regional studies of the Cretaceous in the entire coastal overlap area. The volume will undoubtedly become a standard of reference; and the Department of Geology, Mines, and Water Resources, together with the co-operating companies, is to be complimented on making this information available.

W. C. KRUMBEIN

"Boden- und Klimaverhältnisse in Mittel- und Westeuropa während der Würmeiszeit." By HANS POSER. (*Erdkunde*, Bd. 2, Lfg. 1–3.) Pp. 53–68. Bonn, 1948.

This paper is a co-ordinated condensation of two articles published in *Die Naturwissenschaften* (Bd. 34, pp. 10–18, 232–238, 262–267, 1947). It deals with the time of maximum of the last European glaciation, the Würm-Weichsel (Brandenburg stage). Perennially frozen ground or pergelisol[1] in strict sense and inferred temperature conditions are discussed in the first part. In central and western Europe there is evidence of glacial-age pergelisol in loess and clay wedges, involutions (*Würgeböden*, *Taschenböden*), and climatically induced asymmetric valleys. Loess and clay wedges are superficial features which form nets with large meshes. They represent fillings of ancient ice-wedge nets which develop at very low temperatures. Involutions are excessively disturbed, squeezed, and disrupted structures of originally horizontal beds and are forming today in the pergelisol tundras of Siberia but not in eastern or central Europe. They are caused by pressure exerted through freezing from below and above of the ground thawed in summer, the mollisol. Climatically induced asymmetric valleys, usually with steep west- and south-facing slopes, occur only in the headwater regions (of central Europe). As known from Siberia, such valleys form because slopes facing west and south thaw out sooner and deeper than do slopes facing in other directions and therefore erode more easily.

Data on these three indicators of glacial-age pergelisol in western and central Europe are compiled on a map (fig. 1). On the whole, their northern limit of distribution coincides with the southern boundary of the Weichsel (Brandenburg stage) glaciation. Involutions which occur in places in the region of young drift date from the ice advance. The equatorial border of the pergelisol was mainly dependent on the depth of the snow cover and on the winter temperature and may imply a mean annual temperature of less than −2° C. (28°4 F.; also see Stephen Taber in *Geol. Soc. America Bull.* 54, p. 1506, 1943). It passed from Brittany to the Alps and through northern Yugoslavia and southern Rumania.

The Würm-age timber line, which, like its modern successor, may have coincided with the July-isotherm of 10° C. (50° F.), is taken from the work of Franz Firbas. This line ran somewhat south of the pergelisol limit in France; west, south, and east of the Alps to Czechoslovakia; and then southeastward on the south side of the Carpathians. These boundaries divide central Europe into four main ancient climatic regions: (1) the pergelisol tundra extending from Land's End into Russia; (2) the continental pergelisol forest east of the Alps; (3) the maritime pergelisol-free tundra which forms a narrow strip in France; and (4) the maritime pergelisol-free forest west of the Alps (fig. 6).

At the intersection of the pergelisol and timber lines in northwesternmost Yugoslavia, where the July temperature may have been 10° C. (50° F.) and the annual mean −2° C. (28°4 F.), the temperature is estimated to have been 8°–9° C. (14°4–16°2 F.) lower than at present in July, and 12° C. (21°6 F.) lower in January. In the strip between the two boundaries in France the temperature may have been about 7° C. (12°6 F.) lower in July, and 0°–3° C. (0°–5°4 F.) lower in January.

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[1] The term "pergelisol" was proposed, together with "involutions," "plications," "mollisol," etc., by Kirk Bryan in the *Am. Jour. Sci.*, vol. 244, pp. 622–642, 1946, for perennially frozen ground and to replace "tjäle" and "permafrost." However, as Carl Troll stresses in *Erdkunde*, Bd. 2, p. 6, 1948, intensive frost action is by no means limited to perennially frozen ground but is also produced by winter, short periodic, and nightly freezing of the ground. Troll therefore proposes the additional terms "annual gelisol" and "diurnal gelisol." He states that active annual gelisol is present in the Scandinavian mountains, active diurnal gelisol in many mountains of low and middle latitudes, including those in New England (Ernst Antevs, *Alpine Zone of Mt. Washington Range* [Auburn, Maine: Merrill & Webber Co., 1932], pp. 34, 35, 40, 57, 61).

The second part of the paper deals, first, with the depth of the summer thaw (thickness of the mollisol) during the last glacial maximum and inferred summer climate. The depth of the seasonal thaw is recorded by involutions, plications, or folds (*Stiche*) and, with reservations, by loess and clay wedges. All three features give minimum values. To eliminate topographic influences, only records from relatively level land less than 300 meters (984 feet) in elevation are used. The data are plotted on a map (fig. 2), on which three regions become apparent. The forested Hungarian Basin thawed to a depth of only 1.5–3 meters (4.9–9.8 feet), probably because the vegetation exerted a shielding effect, whereas the tundra from Denmark to the Seine River thawed to 1–3 meters (3.3–9.8 feet), suggesting a long season with positive, even if low, temperature. Central Germany had a thaw depth of only 1 meter, apparently because of strong cooling by the Nordic and Alpine ices.

These temperature conditions and their distribution give indications about the summer pressure, wind, and precipitation. The high atmospheric pressures over the ices may frequently have been connected by a bridge, while low pressures may have prevailed over the Hungarian Basin and northwestern continental Europe (fig. 3). In the region east of the Oder River and the Alps the prevailing winds came from the northeast; in western and central Europe from the southwest. Cyclonic storms must frequently have progressed from the Mediterranean over Hungary and from the Atlantic over France, the Low Countries, and Germany. Precipitation was probably heaviest at the ice border and in central Germany, though this latter region by some students is assumed to have had a cold-dry climate.

Whereas the depth of thaw bears on the ancient summer climate, disruption of already frozen ground by contraction during very low temperatures gives hints on the winter climate. The intensity of the frost tearing can be judged from the depth and relative abundance of frost cracks and clay wedges, these latter requiring lower temperatures. In northwestern continental Europe are found only narrow and, at most, 3-meter (9.8-feet) deep frost fissures, whereas in the interior there are some cracks but predominantly clay wedges, which in central Germany are mostly over 5 meters (16.4 feet) deep (fig. 4). It is concluded that the mean January temperature in the Hungarian Basin was somewhat lower, in central Germany decidedly lower, and in the maritime northwest part of the continent distinctly higher than that at the intersection of the pergelisol and forest limits in northwestern Yugoslavia, where it is calculated to have been about −14° C. (6°8 F.). From these temperature conditions it is, in turn, concluded that the atmospheric pressure was high over most of Europe and that an offshoot of low pressure extended in over northern France and contiguous regions (fig. 5). Southwest winds may have prevailed over western Europe, east and northeast winds in the region east of the Elbe River and the Alps. Except in the west, the pressure conditions may have been very stable, cyclonic storms few, and precipitation small.

The main results of this important study of the ground-frost and climate during the Weichsel-Würm glacial maximum are summed up in a map (fig. 6), purporting to show the then climatic provinces of middle and western Europe, and in an accompanying brief description of the provinces.

Ernst Antevs

Soil Mechanics in Engineering Practice. By Karl Terzaghi and Ralph B. Peck. New York: John Wiley & Sons, Inc., 1948. Pp. xviii+566. $5.50.

The book on *Soil Mechanics in Engineering Practice* by Terzaghi and Peck should prove of general interest to geologists. The science of soil mechanics has developed very rapidly in the last ten years. Since the publication of Terzaghi's work on *Erdbaumechanik* more than twenty years ago, he has been the leader in this field. Thus this book is particularly timely. Engineers in recent years have become increasingly conscious of the need for precise information upon the behavior of soft or loosely consolidated material when it is subjected to loads or stresses. Engineers cannot always pick and choose the places where they are asked to build structures and do their work, so they must do the best they can under the conditions with which they must work.

This book presents a very thorough analysis of present concepts of the principles of soil mechanics and indicates in considerable detail how these principles can be applied effectively to engineering problems. It also includes sections on methods of procuring and testing samples for analysis.

Soil mechanics is useful to the geologist as well as to the engineer. First, it helps in his consulting work on engineering problems, because all the practical applications of soil mechanics to engineering have geologic aspects, of which the geologist should be aware. Such engineering problems as adequacy of the earth to support structures, including both buildings and dams; settlement of the ground after structures are built; stability of slopes of cuts and embankments; earth pressure on retaining walls; strength of earth materials in dams; highway and airport construction; and tunnels in soft ground all concern a geologist at one time or another.

Second, soil mechanics helps the geologist in understanding the mechanics of deformation of rocks, because the basic principles of mechanics apply equally well to hard rocks as to soft rocks. The principles of soil mechanics are particularly useful to the geologist in interpreting the folding and faulting of sequences of interbedded soft and hard rocks and also in studying landslides and hill creep.

Third, compaction of sediments is closely related to the problem of what the soil mechanics engineer calls "consolidation." The engineers have made more progress on compaction than the geologists have, and the latter can profitably read the excellent discussion of this problem in Terzaghi and Peck's book.

The fundamental basis of soil mechanics is the deformation of loosely consolidated earth materials when subjected to stress. Basically, it involves (1) the movement of the water contained in the sediment and (2) the movement of both the water and the solid constituents.

The problem is approached by a study of the hydrodynamic and physical principles involved and by measurements of sediments under field and laboratory conditions. Though the soil mechanics engineer attempts to attack the problem dynamically, his approach thus far has been to a considerable extent one of statics and of comparison with previous experience. The geologist therefore, because he is constantly being compelled to consider dynamics—that is, the changes that take place with time—and because of his knowledge of the habits of sedimentary materials, is in a position to assist the soil mechanics engineer. Terzaghi and Peck point out clearly this relationship between geology and soil mechanics.

The fundamental geologic principles influencing the behavior of water in sediments or in loosely consolidated earth, as this reviewer sees the problem, are: (1) grain-size distribution and shape of particles; (2) nature of mineral constituents, particularly the clays; (3) chemical composition of the water in the sediments, especially with respect to possible base exchange reaction; (4) the stress that is about to be or has been applied to the sediments; and (5) the influence of time, particularly as to past and future rates of change of stress. Of course, it will take a long time before these fundamental principles are understood thoroughly; yet their solution should be an objective of both geology and engineering. Meanwhile, Terzaghi and Peck have given us many fundamental and empirical relationships that can be applied very satisfactorily to engineering and geological practice.

PARKER D. TRASK

Physical Geology. By CHESTER R. LONGWELL, ADOLPH KNOPF, and RICHARD F. FLINT. 3d ed. New York: John Wiley & Sons, 1948. Pp. 602; figs. 365. $5.00.

This is the third edition of a well-known and widely used elementary geology text which appeared in two previous editions (1932 and 1939) under the title *Textbook of Geology*, Part I: *Physical Geology*, by the same authors. The present revision has not involved radical changes in the organization or subject matter; its most striking changes are physical. Instead of the dull maroon or dark green of the previous editions, the present book is bound in the bright crimson and black of the authors' *Outlines of Physical Geology*. The paper on which the book is printed is of far better quality, and on it the photographs reproduce exceedingly well and are also larger and clearer than those in the previous editions. Appropriate new photographs have been added, and there are more of the excellent aerial photographs which distinguished the former editions. The "bleed" arrangement of many of the photographs gives the book a fresh, modern look. Most of the diagrams have been redrawn to give greater clarity or are wholly new.

A notable departure of this edition is the set of 250 original kodachrome slides which were made to accompany it. For years slides have been used to illustrate features and processes in elementary geology; they have usually been dull, black and white (or more often, gray) "lantern slides." In many instances professors, even those well along in years, showed their students the same slides that they had looked at as students in elementary courses—slides often featuring a whiskered geologist in a derby hat or ladies in the costumes of 1900 or thereabouts. The work of conversion of departmental collections to kodachrome slides lagged far behind the popularity of this type of photography, for it is an arduous task. Such a collection as that offered to illustrate this book (price $85.00 with carrying case and manual keying them to particular topics in *Physical Geology*) should be a boon to teachers of elementary geology courses, particularly those in small departments. In a larger department they can well form the nucleus around which a larger collection can be assembled. The slides, available only in the complete set, are made of film described as "unbreakable" and bound into plastic mounts. They can be used in an ordinary kodachrome projector. The photographs were taken by Orlo Childs of the geology staff at the University of Wyoming, who visited classical localities in all sections of the country. An excellent idea that he followed was to take the pictures, wherever possible, in well-known vacation spots, thus drawing on past and future experiences of the average student.

To the book a new chapter has been added, chapter ii, "The Method and Scope of Geologic Studies." In line with modern trends in science education, this chapter contains, first, a discussion (perhaps too brief) of the scientific method, leading to the idea that geology advances through reasoning based on accurate observation of features and processes, with the goal of reading a complete history of each geologic feature and ultimately of earth itself. With a bow to methodology, the writers state that conclusions are significant only as the methods by which they have been reached are fully understood. They believe that, for the elementary student at least, understanding the basic concepts of geology is more important than memorizing a large number of details. (And, incidentally, series of tests on college students have shown that understanding of concepts is far more lasting than knowledge of specific facts, which are fast forgotten.) Following the policy of the other editions, and with this in mind, the writers have placed more of the purely factual background material or "tools" (minerals, rocks, topographic map explanation, and the time scale) in appendixes, with the hope that this material will become the nucleus of laboratory work, inserted where it is needed in the course for understanding the development unfolded in the text.

In a masterly stroke, Dr. Longwell, responsible for chapter ii, illustrates the scientific method by discussing, almost as the student begins the book, a major problem still partly in the "twilight zone." This problem is the relation between continental masses and the deep-sea floors—in other words, an introductory discussion of the broad causes of diastrophism. Not only does this set the stage for the student by showing him an essential problem of geology and of earth's future, but it gives him a large, oriented picture of our planet's processes and problems, bold in outline, on which he can, from time to time, paint in, in their proper places, the concepts and ideas which he learns throughout the book. When he finishes the course, he thus should have made for himself a large, meaningful, integrated picture rather than a series of those perhaps interesting, but unrelated, small pictures which form the collection of many students when they finish textbooks or courses in elementary geology. Even more important from the standpoint of student interest, without which the instructor is powerless to teach effectively, is the fact that a big, important, "dignified" problem has been presented. In many texts and courses on geologic processes the student must spend considerable time, particularly in the early parts of the course, on topics like "work of the wind," "weathering," "work of running water," which, in many cases, fail to catch his interest and respect—because he cannot yet see the magnificent whole into which they will fit. He is likely to contrast it unfavorably with the courses he is taking concurrently—physics, "glamorized" by the introduction of atomic energy; astronomy, where he deals almost immediately with vast distances and majestic vistas of space; chemistry, where he studies the ultimate nature of matter. Dr. Longwell's presentation is skilfully made; the student is given information that he can understand at this point and an incentive for learning the detailed evidence and following the detailed reasoning of the body of the book.

LOU WILLIAMS PAGE

The Religion of the Modern Scientist (Neo-materialism). By S. W. TROMP. Leiden, Netherlands: A. W. Sÿthoff's Uitgeversmaatschappy N.V., 1947. Pp. xxiv+480; figs. 113; 1 table.

The author of this broadly inclusive volume is a geologist, still relatively young, who has spent several years in petroleum exploration in the Dutch East Indies, Egypt, and Turkey and is at present the professor of geology in the Fouad I University in Cairo. His work will therefore be of great interest to all geologists, especially so because more than half the book is devoted to an exposition of geologic principles and an account of earth history. As a matter of fact, the title of the book is likely to be misleading; not that geologists should shy away from it if it were really what one would naturally expect it to be. Actually, there is little in it that can be called "religion," although there is quite a bit of philosophy and much that might serve as a good foundation for an intelligent code of personal behavior.

It is Dr. Tromp's thesis that "the law of action and reaction which determines the absolute causality in the non-living world must be applied also on the living world, being ruled by the same neo-materialistic laws." To demonstrate the correctness of this idea, he presents a digest of all knowledge, running the gamut from the "wonders of the crystalline world" through the minerals and rocks, the structures and history of the earth, the evolution of plants and animals, to mental processes, telepathy, and psychokinesis. With such an ambitious program it is almost inevitable that an occasional erroneous statement crops out here and there. For example, in discussing "experiments with tissue cultures in vitro" on page 199, he describes "vitro" as "a special feeding-fluid."

It is presumably Dr. Tromp's interest in "absolute causality" that impels him to search for "laws" in all aspects of cosmic evolution. What he means by "law" is perhaps illustrated by his "law of varying strike line of oscillations." This is one of his "main laws of mountain building." He defines it as follows: "The strike line of the oscillatory movements in the same region is often subject to variations during the different geological periods." If that is a "law," it denotes a concept of "law" that makes it possible to accept many of his statements pertaining to the deeper problems of human life, such as this one on page 241: "There is no indication that the thinking process of man is ruled by laws fundamentally different from those ruling the crystalline inorganic world."

It is, furthermore, quite logical for him to conclude that "certain symptoms in the hand caused by the distribution of nerve centers might indicate certain phenomena which are going to take place during the future life of that person" and that there may well be something in astrology and the prophecies of mediums. Pursuing that line of thought, he devotes several very interesting pages to a consideration of dowsers and divining rods. Even though one may not agree with his conclusions—and most certainly I do not—the sympathetic analysis that he gives is quite informative and well worth reading.

Even though I cannot agree with Dr. Tromp that the fatalistic determinism of his "neo-materialism" is the religion of the modern scientist, I am glad to recommend most highly his synthesis of the geological sciences. His summary of geologic life-development and of human history from Pleistocene time into the new age of atomic energy, as well as his critique of geologic principles, is cogent and refreshing. It is not seriously marred by the numerous typographical errors and the non-American idiom that frequently diverts the attention of the reader from ideas to words.

KIRTLEY F. MATHER

Mining Geology. By H. E. MCKINSTRY, with sections by STANLEY A. TYLER, E. N. PENNEBAKER, and KENYON E. RICHARD. New York: Prentice-Hall Book Co., Inc., 1948. Pp. 602, plus 57 pp. of appendixes and glossary containing tables of rock classifications, chemical elements, natural functions, etc., and familiar mining and geologic terms; figs. 149; pls. 2; tables 13. $10.00.

This book is a constructive contribution to the progress and development of the application of the science of geology to mining. Here is an orderly, rational, and lucidly written exposition of the ABC's of where and how to look for factual data and how to collect, present, and evaluate them when examining an ore deposit. Moreover, the book explains how these data may be used for guidance in reaching the economic decision as to whether it is an ore deposit from which one or more metals or minerals can be recovered at a profit or just another mineral deposit of no immediate economic importance.

The author divides his book into four parts: (I) "Assembling Geologic Data"; (II) "Geological Principles of Ore-Search and Ore Apprisal"; (III) "Application in Specific Phases of Mining"; and (IV) "Technological Characteristics of Ores." Individual chapter headings include such varied subjects as "Geologic Mapping," "Sampling Ore and Calculating Tonnage," "Drilling," "Guides to Ore-Targets and Loci," "Valuing Mining Properties," "Writing and Reading Reports," "Geological Work at an Operating Mine," "Marketable Forms of Metals and Ores," and a "Glossary of Mining and Geological Terms." The book is well documented with references. As one could expect at a price of $10.00, the book is well illustrated, printed, and bound.

This treatise gives an excellent summary of the responsibilities and duties of the mining geologist. It differs from the many treatises on "mineral deposits," "economic geology," and other textbooks, in that it is devoted wholly to geology. The author does not hesitate to orient and focus his whole effort on the industrial application of the embryo science. Not so much as a paragraph is devoted to questions of origin or other theories whose practical application cannot be foreseen.

The book is primarily concerned with the everyday tasks of the mining geologist—mine-mapping, logging and interpreting diamond drill cores, and analyzing and interpreting geology primarily for the purpose of guiding mine prospecting and exploration. The emphasis is on the search for ore and ore-finding more than on mine-examination work, although, to be sure, one involved strictly in mine examinations should benefit considerably by the geologic approach presented here, for the author approaches this question from a geologist's point of view rather than from a mining engineer's viewpoint. There is already available a long series of excellent books by mining engineers that treat various phases of mine examination, but only Forrester's recent book, *Principles of Field and Mining Geology*, is also written from a geologist's point of view.

While engineering technology (production economics) is unquestionably a vital factor in mine examination, the basic problem may be geologic. It benefits the engineer little to discuss at length in this report the relative economic merits of square-set stoping versus top slicing if his sampling and appraisal of ore reserves is based on wrong geologic assumptions. The selection of the stoping method will, moreover, probably be dictated by geologic factors, such as character and strength of the rock overlying the ore, presence or absence of slips and sheeting, etc. Many concentrating problems are also influenced, if not controlled, by details of the ore occurrence, as the author points out. McKinstry's geologic presentation of the problem is timely.

While the experienced mining geologist who has read widely in his profession may find few new ideas in the book, I am sure he will benefit by the orderly presentation and logical development of the author's thesis. We need more books like this, which show the young geologist how to apply the theory that he has learned in school and how to develop the field application of his profession. I recommend the book to all graduate students in economic geology not only for the sake of its thesis but also as a splendid example of simple, well-organized, lucid, and interesting writing on a technical subject.

Olaf N. Rove[1]

The Iron Resources of California. (*California Division of Mines Bull. 129.*) 1948. Pp. 304.

This book is a valuable résumé of the work done on the iron-ore deposits of California by the United States Geological Survey and the United States Bureau of Mines during the recent world war. Thirteen of the more than one hundred deposits of iron ores known in California are described in detail; an article on the titaniferous magnetites of the western San Gabriel Mountains in Los Angeles County is included, as is also a summary of the present iron-ore situation in California and a short section describing the results of the diamond drilling conducted by the Bureau of Mines on five of the properties. The thirteen deposits described were selected because it was thought that they were the largest and had the greatest possibilities for commercial development. The deposits are the Lava Bed district, the Silver Lake district, Old

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[1] Published with the permission of the Director, U.S. Geological Survey.

Dad Mountain, Cave Canyon district, the Vulcan deposits, the Iron Hat district, Ship Mountains, and the Kingston Range, all in San Bernardino County; the Minarets deposit along the crest of the Sierra Nevada Range in Madera County; the Shasta and California deposit and the Hirz Mountain deposit in Shasta County; and the deposits at Lake Hawley and Spencer Lakes in Sierra County. The five deposits drilled by the Bureau of Mines are Eagle Mountains, Silver Lake, Lava Bed, Shasta and California, and Minarets.

The total iron-ore reserves of California are estimated at between 100,000,000 and 150,000,000 long tons, of which about half is not now commercial, either because of the long distance to a market or because of the small size of the deposits. The Eagle Mountains area contains about a third of the total reserves and is the only deposit in California capable of supporting a modern blast furnace for a period of ten years or more. Reserves of all the other deposits examined, except that at Kingston Mountain, are given and range from about 6,550,000 long tons at Silver Lake to about 80,000 long tons of magnetite at Ship Mountain. The Vulcan deposit furnished 2,643,000 tons of ore to the Fontana steel mills during the period 1942–1944. Several of the other deposits have been worked in a small way for special-purpose ores.

The iron-ore deposits, except for those at Minarets, at Ship Mountain, and at Lake Hawley and Spencer Lakes, are typical contact metamorphic replacement bodies in limestones and dolomites. The ores in all the deposits are magnetite, except at Eagle Mountains, Kingston Range, and Ship Mountain, where the ores are mixtures of hematite and magnetite. At Minarets the magnetite ores replace meta-andesite and are not related to any known intrusive mass. The small deposits around Lake Hawley and Spencer Lakes are described as replacement bodies in clastic sediments, dolomites, tuffs, and lamprophyric dikes. They are genetically related to an intrusive mass of metadiorite and its associated dikes. The ores at Ship Mountain form lenses in brecciated igneous and metamorphic rocks. Gypsum is present in the outcrops of several of the deposits, and in parts of several ore bodies the sulfur (pyrite) content is high for direct smelting. Phosphorus is low in all the analyses given.

On the surface the Kingston Range deposits appear to be among the largest and most promising of California; but, from diamond drilling done in 1924 and from later detailed geologic mapping, it is concluded that the ores lie in the overthrust block of a large thrust fault and that they do not continue into the underlying rocks.

The titaniferous magnetite bodies of the San Gabriel Mountains are in gabbro and anorthosite. The ores range from low grade to nearly solid titaniferous magnetite. In part of the rock the ilmenite and magnetite are minor accessory minerals, but they increase gradually in amount until some bodies are composed entirely of an intergrowth of magnetite and ilmenite. Thirty-six deposits were mapped and described. The best of the bodies contains about 250,000 tons of mineral, carrying 11–24 per cent of TiO_2. Details of a few placer deposits, found in the streams below the igneous masses, are given.

This book is mainly descriptive, but it brings together a great deal of miscellaneous information about the iron-ore deposits of California. It is a worth-while contribution and will be of special interest to anybody who desires to obtain impartial information about the character and the reserves of California iron-ore deposits.

CHARLES F. PARK, JR.

COMMUNICATIONS AND ANNOUNCEMENTS

GRANTS-IN-AID AVAILABLE FOR SCIENTIFIC RESEARCH

The American Academy of Arts and Sciences announces that income from its Permanent Science Fund will be disbursed as grants-in-aid in support of research projects in the fields of scientific business management, manufacture and commerce, engineering, psychology, education, economics and sociology, mathematics, physics, chemistry, astronomy, geology, geography, zoölogy, botany, anthropology, history and philology, medicine, surgery, agriculture, or any other science of any nature or description.

Applications for grants-in-aid are receivable on multiple forms, which will be supplied upon request to the chairman of the Permanent Science Fund Committee and are considered by the Committee on March 1 and October 1.

Requests for further information about conditions governing the grants should be addressed to: Hudson Hoagland, Chairman, Permanent Science Fund Committee, Worcester Foundation for Experimental Biology, 222 Maple Avenue, Shrewsbury, Massachusetts.

ERRATA

In "The Geomorphic History of the Carlsbad Caverns Area, New Mexico," by Leland Horberg, p. 472, legend for plate 1: reverse *B* and *C*.

In "The Ogallala Formation West of the Llano Estacado," by J Harlen Bretz and Leland Horberg, p. 492, legend for plate 1: reverse *B* and *C*.

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